



SOILS

THEIR ORIGIN, CONSTITUTION, AND CLASSIFICATION

AN INTRODUCTION TO PEDOLOGY



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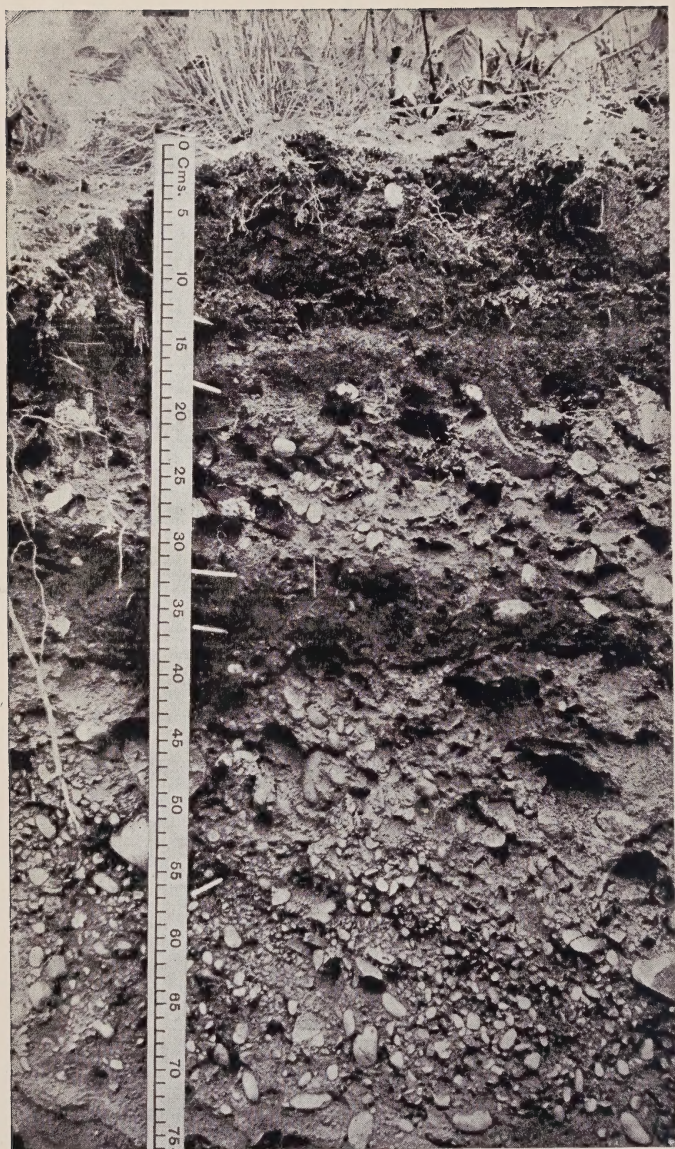
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[Photograph by W. Morley Davies,
Podsol profile at Goldstone, Salop.

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SOILS

Their Origin,
Constitution, and Classification

AN INTRODUCTION TO PEDOLOGY

BY

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TO
CURTIS FLETCHER MARBUT

“ Habent enim rationem cum terra, quæ numquam recusat imperium nec umquam sine usura reddit quod accepit, sed alias minore, plerumque maiore cum fænore. *Quamquam me quidem non fructus modo sed etiam ipsius terræ vis ac natura delectat.*”

CICERO. *De Senectute.*

PREFACE.

THE student of *Don Quixote* may remember the criticism of Cervantes on one of his own works: "*He propounds something and concludes nothing.*" On looking over my proofs, I am aware of the aptness of this judgement to the present work. My aim has been to give, within the compass of a book of moderate size, a general view of pedology—the study of soils. Recent developments, however, have so increased the scope of the subject that only the broadest treatment is possible if an adequate perspective is to be given within the limits proposed.

The first section of the book is occupied with the origin, constitution, and properties of soils, and, since the standpoint adopted is the exhibition of soils in their natural relationships, I have exercised some selection by dealing mainly with those topics which appear necessary for this purpose. The next section is devoted to the description, with illustrative examples, of the chief soil groups of the world. This is followed by a discussion of the problem of classification, and an account, given with due reserve, of the geographical distribution of soils. In the remainder of the book, I have dealt with soil surveys and soil analysis, and have concluded with a brief discussion of the inter-relationships of soils, plant growth, and agriculture.

I have written primarily for those who are interested in the soil as an object of study in itself, and secondarily for those whose interest lies in its economic or geographical significance; but it may be hoped that the book will prove of value to botanists because of the importance of the soil in

ecology, and to geologists because of the part played by the soil in the weathering cycle.

Since the subject is here treated in its purely philosophical aspect, I have avoided the consideration of certain subjects, such as lime requirements and exchange acidity, which, though occupying important places in the literature, are definitely problems of applied pedology. I have also refrained from the discussion of those enquiries, belonging rather to physical chemistry or physics than to pedology, in which the soil appears merely as a research material. The omission of these subjects implies no adverse reflection on their value or on the ingenuity of the investigations by which they have been illustrated: it is imposed by the necessity of giving, within the available space, the fullest possible picture of soils in their natural relationships.

In developing my account, I have referred only to such investigations as appear necessary to illustrate the main arguments. The bibliography is not, therefore, as full as might be desired by the research worker. The references given at the end of each chapter will, however, lead the reader to the sources of further information. The frequency with which I have used examples from the work of my own laboratory will not, I trust, be held to betoken an undue estimate of its importance. I have made free use of this work because it is most familiar and lies most readily to hand for purposes of illustration.

My cordial thanks are due to those authors who have so kindly allowed me to make use of material from their published works or private communications. Special acknowledgements are also due to Messrs. Paul Parey, Berlin; Gebr. Borntraeger, Berlin; The Macmillan Company, New York; The Williams and Wilkins Company, Baltimore (*Soil Science*); and The Cambridge University Press (*Journal of Agricultural Science*) for permission to reproduce tables or diagrams from their publications.

I am particularly indebted to Mr. G. H. Freeman (Messrs. Thos. Murby & Co.), whose sympathetic co-operation has made the author's share of the work of publication a pleasant and interesting task.

Finally, I wish to record my indebtedness to the colleagues, past and present, who have worked with me at Bangor. They have helped in many tangible ways, but most of all by their loyal support throughout many happy years of study.

G. W. ROBINSON.

UNIVERSITY COLLEGE,
BANGOR,
Easter, 1932.

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CHAPTER I.

INTRODUCTORY.

THE study of the soil, from its close connexion with the important art of agriculture, has occupied, until recently, an intermediate position between pure and applied science. Most of the earlier investigations on soils have been directed towards the solution of problems arising from practical agriculture, and this circumstance has introduced complications both in the statement of the subject and also in the methods of research employed. The investigator of the relationships between the growing plant and its environment is often in danger of failing to secure an adequate view either of the soil itself or of the subject of plant nutrition.

Students of the soil are, therefore, under a debt to such leaders as Docuchaiev and Glinka in Russia, Ramann in Germany, and Hilgard in America, who have succeeded in creating a body of effort directed towards the investigation of the soil as a definite branch of natural philosophy. For this study, the name *pedology** has been proposed. In spite of the reluctance in some quarters to accept this reasonable name for the study of soils, it can hardly be maintained that there is a satisfactory alternative.

Pedology is the study of the soil from the standpoint and by the methods of pure science. In one aspect, it is a branch of geology, since it is concerned with the superficial strata of the regolith†, in which most of the important processes of weathering and denudation are fulfilled. In

**Pēdology*, from *πέδον* = ground or earth. It should not be confused with *pēdology* (or *pædology*), a term sometimes used in America for the study of children.

†*Rēgolith* (or better, *rhēgolith*), the fragmental unconsolidated débris, mantling the rocks of the earth's crust.

another aspect it is the study of the physical and chemical properties of a complicated colloidal system. In a third aspect it is the study of a complex flora and fauna in relationship to soil environment. These three aspects are closely interdependent, for a complete knowledge of the origins and metamorphoses of soils involves an understanding of soil constitution, which itself is closely related to soil microbiology.

The adoption of the view that pedology should be regarded as an independent branch of study and treated as a pure science does not imply any depreciation of the practical significance of the subject. On the contrary, it may be submitted that, by elucidating the origin and constitution of the soil, and by developing a philosophical system of classification, surer progress can be made than by trammeling investigation with economic and practical considerations, which so frequently have the effect of obscuring otherwise straightforward problems. The manner in which pure science has contributed to practical life during the past century affords ample justification for this view of the relative importance of the pure and the applied standpoints in the study of natural phenomena. As the discoveries of the pure chemist and the pure physicist have found applications in daily life, so we may expect that, when the constitution of the soil is more perfectly understood, applications to practical agriculture will also be forthcoming.

In a relation of the origin, constitution, and classification of soils, the familiarity of the subject might almost be held to render a definition superfluous. Yet, viewing the soil as the most superficial layer of the regolith, the student is at once faced with the difficulty of deciding where the study of the soil ends and where the domain of the geologist begins.

Where a thin layer of soil lies directly on the rock from which it has originated, there is every justification for regarding all horizons down to the unaltered rock as within the province of the pedologist. The position is less clear

when we consider soils derived from deep unconsolidated deposits such as alluvium, loess, or glacial drift. To restrict the study of the soil to the upper layers in which plant roots ordinarily fulfil their activities would be to exclude horizons which clearly fall within the sphere of action of the *pedogenic* (soil-forming) processes. To define the soil as those horizons affected by the pedogenic processes would be, of course, to state an identity. Nevertheless, a study of deep sections suggests that below a certain depth, generally not more than a few feet, the strata are purely geological, and only indirectly affected by the pedogenic processes which operate in the upper horizons. On the whole it seems preferable, in the present state of our subject, to decline a definition and merely to indicate that the range of interest of the soil student is not confined to the depth utilised by plant roots and subjected to tillage by the cultivator. By adopting this position, the pedologist claims the right to study the pedogenic processes in their widest significance. The domain of pedology can only be defined by the natural limits of enquiry and may come to engross a considerable portion of dynamic geology. Indeed, studies of contemporary pedogenic processes may be expected to throw light on many of the problems of sedimentary petrology.

Considered in quantitative relationship to the lithosphere, the regolith is of trifling thickness, whilst the soil, the superficial stratum of the regolith, is even more insignificant. Yet the importance of the soil in dynamic geology can hardly be over-estimated, for it is the seat of the most active operation of epigene weathering and is also that part of the earth's crust immediately subject to denudation. The soil is thus an essential phase in the weathering cycle, and its study is a branch of geology in its broadest sense.

Soil consists essentially of (a) mineral matter, which has originated from rock by the action of a series of weathering processes, (b) organic matter, which has originated from the residues of natural vegetation and organic manures, (c) soil

moisture, containing substances in colloidal or in true solution, and (d) soil air. The mineral matter includes particles of varying grade from stones and gravel down to sub-microscopic particles of clay. In it may be recognised particles of rock-forming minerals, and material which is either amorphous or else consists of minerals whose crystalline character can only be revealed by X-ray analysis. The latter material is called the colloidal clay, or clay fraction, and confers on soils and clays those properties, such as plasticity and cohesiveness, which distinguish them from mechanically comminuted rock. Whilst the physical disintegration of rocks can account for the recognisable rock and mineral fragments in soils, the clay fraction results from processes of chemical weathering, mainly silicate hydrolysis. We shall, in a later chapter, discuss this process with more minute attention.

The organic matter of soils is present mainly in the form of dark-coloured amorphous material, which we may designate as humified organic matter or, simply, *humus*. It has originated naturally from the residues of vegetation, but in cultivated soils organic manures have also contributed. Fragments of undecomposed plant-remains, leaves, stems, roots, and seeds, can also be recognised, but from 60 to 90% of the organic matter of soils is in the humified state. Organic matter added to the soil may be either oxidised to carbon dioxide or, under conditions which we shall discuss later, humified. The amount of organic matter in a soil is thus governed by the extent to which it is enriched by plant residues, and also by the relative preponderance of the oxidative and humifying processes. Whilst desert soils may contain fractional percentages of organic matter, the prevalence of the humifying processes in humid climates may result in the production of peat soils consisting almost entirely of organic matter. In ordinary English soils the proportion of organic matter varies from 3 to 10%.

The most important advance during the past generation has been the recognition of the soil profile as the unit of

study. To an earlier generation of investigators, the individual soil meant a sample of soil, generally from the superficial layer occupied by the roots of plants. We now know that an adequate conception of the soil can only be obtained from a study of all the horizons down to the parent material. And whilst for some purposes it is sufficient to study the properties of samples taken from the recognisable horizons of a profile, a complete picture can only be obtained when the laboratory studies are supplemented by observations on natural profiles in the field.

The investigations of recent years have been for the most part directed towards the elucidation of the soil viewed as a static system. Much of this work belongs as much to colloid chemistry and physics as to pedology. But, eventually, the pedologist must go into the field and study his material under natural conditions.

The past years have been fruitful in perfecting the instruments of research. In the writer's opinion, the centre of interest must now shift more and more from the laboratory to the field. It is therefore to be expected that the future contributions to pedology will tend to be mainly descriptions of new types of soil, and studies of the chemical, physical, and biological processes which take place under field conditions.

The soil profile—the succession of horizons down the parent material—gives expression to all the processes of soil formation, including physical and chemical weathering, and the vertical translocation of soil constituents by water movements. The character of these processes, and hence the character of the soil profile, is mainly governed by climate. Climate also controls the type of natural vegetation, which in its turn reacts on profile development. But the parent geological material also plays a part, and under certain conditions may prove the most important factor in differentiating soils from each other.

Fully developed soil profiles are not everywhere in evidence. In some cases the soil profile is immature, as, for

example, in soils developed on recent alluvial or æolian deposits, where time has not sufficed for the full expression of the pedogenic processes. In other cases, the profile may have been truncated by erosion, whereby some or all of the horizons have been removed, the present-day vegetation being established on a lower horizon of the original profile. The natural profile may also be disturbed by cultivation, which results in a confounding of all horizons within the reach of the implements of tillage.

Since the character of the soil profile is affected by the mode of action of the pedogenic processes and by the nature of the parent geological material on which these processes have operated, it will be readily understood that the possibility of variety in the final result is very great. Yet we shall see in the sequel that, in the main, the major classification of soils is closely related to climate, whilst geology is chiefly of significance in transition regions and in the later stages of classification. Topography also enters as a factor and may profoundly modify the operation of the regional climatic influences.

In writing on the subject of soils, some difficulty is experienced in securing a satisfactory order of presentation. Thus, it is difficult to discuss the pedogenic processes without reference to the colloidal complex, and this, in its turn, involves the study of organic matter and of base exchange. The inter-relationships of different branches of the subject have sometimes entailed a certain amount of repetition and the appearance of the same topic at more than one point in the narrative. The reader who is making his first acquaintance with the subject may therefore find it advantageous to re-read some of the earlier chapters in the light of later chapters.

The chapter on Soil Classification has been placed after the chapters in which the principal soil groups are described. This seemed advisable because the problems involved can only be appreciated when the reader has gained some knowledge of the material which it is proposed to classify.

GENERAL BIBLIOGRAPHY.

The following books may be recommended to those who wish to study the subject in greater detail:—

- E. W. HILGARD. *Soils*. The Macmillan Company, New York. Of classical interest as the work of the founder of modern pedology.
- E. RAMANN. *Bodenkunde*. Julius Springer, Berlin. Also a classical work, it forms a bridge between Hilgard and contemporary authors.
- E. J. RUSSELL. *Soil Conditions and Plant Growth*. Longmans, Green & Co., London. An authoritative text-book with an exceedingly copious bibliography.
- A. STEBUTT. *Lehrbuch der Allgemeinen Bodenkunde*. Borntraeger, Berlin. An excellent presentation of the subject from the purely philosophical standpoint.
- K. D. GLINKA. *Die Typen der Bodenbildung*. Borntraeger, Berlin. "The Great Soil Groups of the World and Their Development"—English translation of the above by C. F. Marbut. Edwards Bros., Ann Arbor, Mich., U.S.A. T. Murby & Co., London.
- E. BLANCK, ET. AL. *Handbuch der Bodenlehre*. Ten Volumes. Julius Springer, Berlin. A standard work to which numerous authorities have contributed.
- P. VAGELER. *Grundriss der tropischen und subtropischen Bodenkunde*. Verlagsgesellschaft für Ackerbau, Berlin. A suggestive introduction to the study of tropical soils.
- B. A. KEEN. *The Physical Properties of the Soil*. Longmans, Green & Co., London.
- S. A. WAKSMAN. *The Principles of Soil Microbiology*. Baillière, Tindall & Cox, London.

The following journals are devoted wholly or partly to publications on the study of soils:—

- Journal of Agricultural Science*. Cambridge University Press, England. (Partly Soils.)
- Soil Science*. The Williams and Wilkins Co., Baltimore, U.S.A.
- Zeitschrift für Pflanzenernährung, Düngung, und Bodenkunde*. A. Wissenschaftliche Teil. B. Angewandte Teil. Verlag Chemie, Berlin.
- Soil Research, Bodenkundliche Forschungen, Recherches sur le Sol*. Published by the International Society of Soil Science, under the Editorship of Prof. Dr. F. Schucht. Berlin N.4, Invalidenstrasse 42.
- Journal of the American Society of Agronomy*. Published by the Society. Geneva, N.Y., U.S.A. (Partly Soils.)
- Internationale Mitteilungen für Bodenkunde*, 1911-1924.

CHAPTER II.

GENERAL VIEW OF THE CONSTITUTION OF THE SOIL.

PROXIMATE CONSTITUENTS.

THE most superficial consideration of soils will reveal the immense variety of their composition and structure; but there are, nevertheless, certain groups of constituents present in all soils, which impart a unity to their study. Briefly, these groups are (1) *mineral matter*; (2) *organic matter*; (3) *soil moisture*; and (4) *soil air*. The last two are, strictly speaking, dependent variables and, in amount and character, controlled by the mineral and organic matter. We may thus conceive of the soil as a three phase system with an indefinite number of components.

The solid phase is composed of the mineral matter and the organic matter. The liquid phase is in equilibrium both with the solid phase—the mineral and organic matter—and the gaseous phase—the soil air. But the equilibrium is continually shifting owing to the variations in temperature and in water content, the drain of nutrients by plant roots, and the activities of micro-organisms. With these variations are associated corresponding changes in the composition of the gaseous phase—the soil air. It is obvious, then, that the solid phase of the soil offers a surer basis of study than the rapidly changing and dependent liquid and gaseous phases. Indeed, for many purposes, the study of the soil is synonymous with the study of soil mineral matter and soil organic matter in the dried laboratory sample. But, when

it is necessary to link up the study of the soil with that of the growing plant, a more complete knowledge of the conditions in the soil solution and the soil air becomes essential.

Considering, then, soil as it appears in the laboratory, we note that it is a mixture of mineral material, which consists of the products of physical and chemical weathering of rocks, and organic matter, which consists, similarly, of the more or less decomposed residues of plant materials and, to a less extent, of animal remains and excreta.

It is not possible by any known means to effect an exact separation of the mineral and organic material of soil, and this constitutes a serious obstacle to its study. By boiling with hydrogen peroxide (15-30%), however, practically the whole of the organic matter is either oxidised to carbon dioxide and water or else decomposed to form soluble products which can be removed by filtration. It is thus possible to obtain the mineral portion of the soil practically without admixture. Unfortunately, no method is known, or seems likely to be discovered, whereby the organic matter can be isolated unaltered, and it is thus necessary to study it either in predominantly organic soils or else in the presence of a great excess of mineral matter.

The mineral matter of soil consists of particles varying in size from stones and gravel down to the submicroscopic particles of colloidal clay. Where a complete range is present, it will be seen that the larger particles consist of rock fragments, the intermediate particles of rock-forming minerals, and the smallest particles of the secondary products of chemical weathering. Very approximately we may say that, in ordinary European soils, fragments above 2 mm. in diameter are mainly rock fragments; particles from 2 mm. to 0.002 mm. in diameter are mainly original mineral fragments, whilst the material below 0.002 mm. in diameter consists principally of the products of chemical weathering.

In certain tropical soils, material of secondary origin may occur among the coarser fractions. Indeed, in highly developed laterites, the surface layer may consist almost

entirely of coarse material, which is of secondary origin and contains comparatively small amounts of original minerals.

MECHANICAL COMPOSITION OF SOILS.

The mineral portion of soil consists of particles varying in size from stones to clay. A general knowledge of the *mechanical composition* of the soil, *i.e.*, the relative proportions of the different sized particles, is of the highest importance both for the purpose of characterising the soil and also because it affords an indication of its physical properties, more particularly in respect of water relationships. *Mechanical analysis*, the series of operations whereby mechanical composition is determined, is the most important laboratory examination to which soil is submitted.

It is not our purpose, in the present chapter, to enter into the details of mechanical analysis as an analytical procedure; yet some allusion to the laboratory aspect of this subject is necessary in order to understand the significance of the results obtained.

In the first place, it is not generally possible to include the larger particles, gravel and stones, in the statement of mechanical analysis. Their inclusion would introduce difficulties both in sampling and also in the actual laboratory procedure. It is therefore customary to make an arbitrary separation of the coarser particles by sieving. Although not universally adopted, the usual convention is to reject all material retained by a 2 mm. round-hole sieve. If desired, the proportion of this coarser material can be determined and expressed separately. The mechanical analysis, then, refers to material whose particles are all less than 2 mm. in diameter. As stated above, the material of coarser grade consists principally of unweathered rock fragments.

Two questions at once confront the analyst. In the first place, it is necessary to decide upon the number of fractions to be distinguished and to define these fractions by particle size. Much confusion has hitherto existed on this point owing to the use of different scales in different

countries. It has now been decided by the International Society of Soil Science that the following fractions shall be distinguished :—

Fractions			Diameter limits mm.
Coarse Sand	...	...	2.0-0.2
Fine Sand	...	...	0.2-0.02
Silt	...	...	0.02-0.002
Clay	...	...	<0.002

In addition to its simplicity, the above scale has the advantage that the successive particle size limits are at equal logarithmic intervals.

Secondly, it is necessary to decide as to the basis on which the results of mechanical analysis shall be expressed. Here again there is a choice of procedure. The results may be expressed either on the air-dry soil, which contains a few per cent. of hygroscopic moisture, or on the soil dried at 100-105°C. There is also a variety of practice with regard to the inclusion of organic matter and calcium carbonate in the fractions, and as to whether the figures given shall refer to ignited or oven-dry weights. In the method now adopted internationally, calcium carbonate and organic matter are removed by preliminary treatment, and the results refer to the carbonate-free mineral material, dried at 100-105°C. A mechanical analysis thus shows the percentages of the four fractions, and of the calcium carbonate, if present; the difference from 100 is principally the organic matter, of which, however, a separate determination must be made, since it is here loaded with all the errors of the mechanical analysis. It should be added that in the method at present used, whilst organic matter is to a large extent removed in the preliminary treatment, the removal is not complete, so that actually the fractions are slightly magnified by the inclusion of such organic matter as survives the treatment used for this purpose, namely, digestion with 6% hydrogen peroxide. The error thus introduced is relatively unimportant for ordinary purposes.

The validity of a mechanical analysis depends on the thoroughness with which the soil is resolved into its constituent particles by the dispersive treatment which precedes the actual mechanical analysis. For this reason a considerable amount of effort has been directed towards discovering the most efficient method of preliminary treatment. The aim of this preliminary treatment is to break down compound aggregates and to bring the soil to its *prime particle structure*. For details of the methods used the reader is referred to the Appendix. Many of the earlier analyses are untrustworthy because complete dispersion was not effected, with the result that a proportion of the material analysed consisted of compound aggregates and not of ultimate particles.

Since the organic matter in soil does not generally exceed 10%, the mechanical analysis will usually account for 90% or more of the soil unless notable proportions of calcium carbonate are present. Dispersion having been effected by preliminary treatment, the coarsest fraction, namely, that consisting of particles between 0.2 mm. and 2.0 mm. in diameter is separated by means of sieving, using a 70 mesh brass wire-cloth sieve, whose apertures are slightly less than 0.2 mm. square.

The finer fractions are determined by the application of a principle enunciated in Stokes' Law, namely, that the limiting velocity of a particle falling in a fluid is proportional to the square of its diameter. The separations can be effected by repeated decantation of the soil suspension, choosing appropriate settling times and depths, *e.g.*, 8 hours and 10 cm. for clay. After each decantation of the supernatant suspension, water is again added; the suspension is thoroughly stirred, again adjusted to the mark, and allowed to settle again. The process of stirring up, settling, and decantation is repeated until, at the end of the time chosen, no more material remains in suspension. The material removed by decantation may be collected and weighed, or it may be estimated by the loss in weight of the material from which it has been removed.

Another method¹ used in mechanical analysis depends on a determination of the concentration of a settling column of soil suspension after a given time at a given depth. For example, to determine the clay, a sample of suspension may be taken at 10 cm. from the surface after settling for 8 hours. The sampling is performed by means of a pipette introduced into the suspension and the concentration found, expressed as a percentage of the original concentration, gives the percentage of clay, *i.e.*, of material having a settling velocity of less than 10 cm. in 8 hours. Experimental details of the method will be found in the Appendix.

The principles underlying mechanical analysis are discussed here in order to show that the lower particle sizes used in mechanical analysis are purely inferential. Apart from the coarsest fraction, each fraction is defined in terms of a settling velocity; the initial assumption, based on the work of A. Atterberg, being that a particle diameter of 0.002 mm. corresponds with a settling velocity of 10 cm. in 8 hours at 20°C. In the writer's opinion, the use of particle diameter is open to criticism, and may be misleading. Only if it were possible to make the separations throughout by sieving would the use of particle size be justifiable. On the other hand, whilst definition of the fine particle sizes in terms of settling velocity is more logical, the fact still remains that the coarsest fraction can only be separated by sieving. It is, however, possible, by application of Stokes' Law, to define the coarser particles in terms of settling velocity or its logarithm. The correspondence of the two methods of definition is shown in the following table:—

Fraction	Diameter Limits	Settling Velocity (V)	Log V
Coarse Sand	2.0-0.2 mm.	347.6 cm./sec.	2.5412
Fine Sand ...	0.2-0.02 „	3.476 „	0.5412
Silt ...	0.02-0.002 „	0.03476 „	2.5412
Clay ...	<0.002 „	0.0003476 „	4.5412

¹ For numbered references see Bibliography at end of chapter.

It will be seen that, whether the fractions are expressed as diameter limits or velocity limits, assumptions are unavoidable; but, in view of the importance of the finer fractions, and the frequent use of the principle of subsidence to determine still finer fractions, the use of settling velocity or its logarithm appears to the writer to afford a more satisfactory basis.

In Table I are given examples of the mechanical analyses of some typical soils.

TABLE I. — MECHANICAL ANALYSES OF SOME TYPICAL SOILS. PERCENTAGES OF SOIL DRIED AT 105°C. FRACTIONS DRIED AT 105°C.

Fraction	Sandy Loam *Shropshire	Light Loam Anglesey	Heavy Loam Anglesey	Rendzina clay Czecho- slovakia	Clay East Africa
Coarse Sand ...	66.6	27.1	13.6	0.9	1.0
Fine Sand ...	17.8	30.3	17.4	7.1	4.1
Silt ...	5.6	20.2	24.7	21.4	7.9
Clay* ...	8.5	19.3	35.1	65.8	82.8
CaCO ₃ ...	nil	nil	nil	2.0	nil
Org. Matter, etc. dissolved in pre-treatment	1.5	3.1	9.2	2.8	4.2
Total ...	100	100	100	100	100
* Including sesquioxides dissolved in pre-treatment.					
	0.4	1.5	1.4	0.7	0.4

MECHANICAL COMPOSITION AS A CONTINUOUS FUNCTION OF PARTICLE SIZE.

The expression of the mechanical composition of a soil as percentages of a limited number of fractions is open to objection on account of the necessity for making arbitrary particle size or settling velocity divisions. Such divisions could only be justified, apart from their convenience, if it could be shown that each limit corresponds with a marked

and discontinuous change in properties. For example, the separation of silt from fine sand at 0.02 mm. diameter would be completely justified if it could be proved that there is an essential difference in constitution and properties between particles of 0.021 mm. and particles of 0.019 mm. diameter. Such a proof cannot be furnished, for very little difference exists. The separation of clay at 0.002 mm. has more justification. At or about this particle size there is a marked change in constitution and properties. Whilst the material above 0.002 mm. consists mainly of rock fragments and chemically unweathered mineral particles, the material below this size consists mainly of the colloidal products of chemical weathering and is the chemically reactive portion of the soil. Whether 0.002 mm. is the most satisfactory limit remains to be ascertained. There is some reason for thinking that a more satisfactory separation might be obtained by the choice of a lower limit, say 0.0006 mm. It should always be remembered, however, that the actual basis of separation is the settling velocity.

On account of the obvious drawbacks to arbitrary size limits, much ingenuity has been devoted to devising methods whereby the mechanical composition of soils can be expressed as a continuous function of particle size. S. Odén² first attempted this by means of an apparatus in which the mechanical analysis of a soil or clay is automatically recorded by measuring the accumulation of falling sediment on a balance pan. G. Wiegner³ also devised an apparatus in which the change in density of a column of sedimenting suspension can be continuously recorded.

MECHANICAL COMPOSITION CURVES.

G. W. Robinson⁴, using the pipette method of analysis, depending on the change in concentration of a settling suspension with time, has shown that, if the mechanical composition of a soil or clay be expressed by means of a curve connecting summation percentages with the logarithm of particle size, smooth curves are obtained which exhibit cer-

tain regularities. The significance of such curves will be appreciated by reference to *Figs. 1 and 2*, in which some typical mechanical composition curves are shown. It will

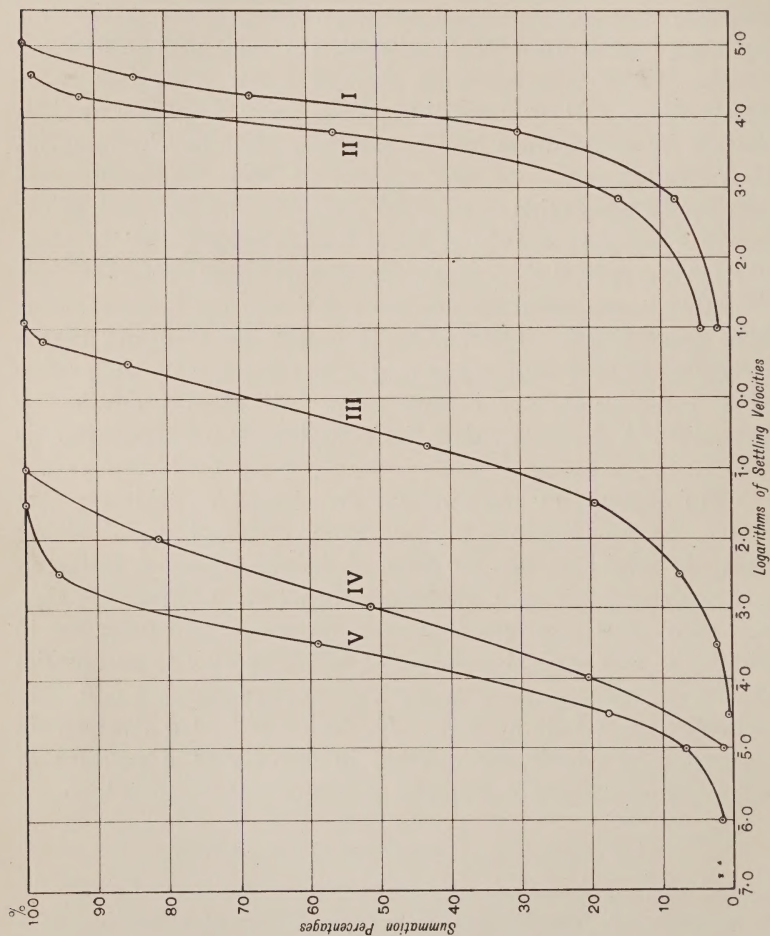


FIG. 1.—Mechanical composition curves of physically comminuted materials.

be remembered that in an ordinary mechanical analysis the fractions reported exclude calcium carbonate and organic matter. For convenience of representation the data on

which the curves are based have been calculated as percentages of the humus-free mineral soil and thus add up to 100 per cent.

The curves shown in *Fig. 1* relate to materials in which, with the exception of kaolin, comminution has been entirely physical.

Curves I and II have been constructed from some data presented by W. Shaw⁵ for the percentage of material passing different sized screens in the preparation of road metal.

Curve III relates to a sample of quartz powder obtained by passing crystalline quartz through a Weatherhead mortar.

Curve IV relates to a commercial sample of kaolin.

Curve V relates to an air-floated powder, Fullersite, obtained from the grinding of slate waste in the Penrhyn Quarries, near Bangor.

The most remarkable feature of the curves in *Fig. 1* is their smooth character, which has been verified also for other mechanically comminuted materials. The general form of the curves is sigmoid, but the upper limb is generally less pronounced than the lower limb. Over the greater part of their range the curves approximate to linear form. The slope of a curve at any point is a measure of the frequency of the fraction represented by the corresponding settling velocity. At the point of inflection, the curve is steepest and this corresponds with what may be termed the *modal fraction*. In other words, if the fractions be chosen at equal logarithmic intervals, the largest fraction is shown by the steepest portion of the curve.

It is of interest to note the essential similarity between Curves I and II for road metal with particles up to about 10 cm. in diameter, and Curve V for slate dust in which the coarsest particles have a diameter of about 0.02 mm.

The curves in *Fig. 2* refer to natural soils and clays, in which physical and chemical agencies have contributed in varying degrees to their comminution. The materials represented are the following :—

- Curve I. Sandy red loam, Tanganyika.
 Curve II. Shale loam, Caernarvonshire.
 Curve III. Clay loam subsoil, Hertfordshire.
 Curve IV. Carboniferous subsoil clay, Flintshire.
 Curve V. Clay fraction from æolian soil, Sudan.
 Curve VI. Red clay, Kenya.

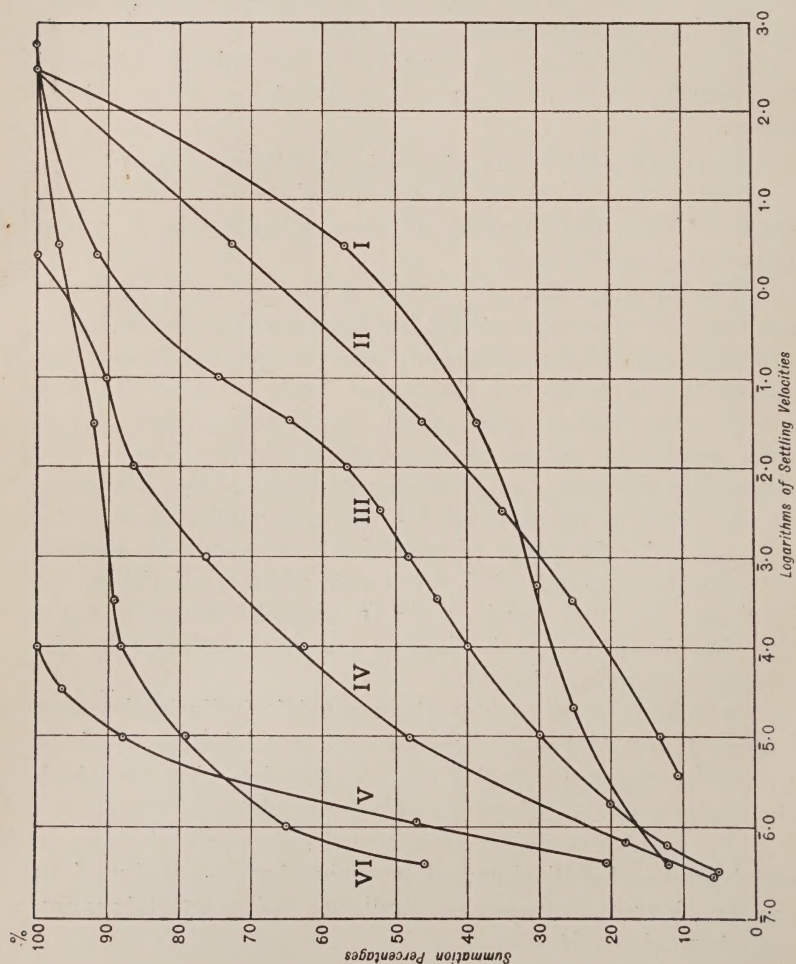


FIG. 2.—Mechanical composition curves of soils.

The smooth character of the curves is again evident, but, except in Curve II., there are marked differences from the curves in *Fig. 1*. The lower limb of the sigmoid appears to be suppressed and the steepest part of the curve is at the lower end of the range. The limit is, of course, obtained by extrapolation, and is therefore conjectural, but appears to lie between $\log V$ 7.0 and 7.5. The latter limit would correspond with a particle diameter of about 0.00006 mm.; but apart from the doubtful validity of Stokes' Law for such small settling velocities, there are disturbing aqueous sheath effects to which reference will be made at a later point. It is therefore unsafe to take the diameter given as actually representing a lower limit of particle size.

Considering the curves in detail, it is to be noted that Curve V relates to a material from which mechanically comminuted particles are entirely absent. Curves I, III, IV, and VI, indicate varying and decreasing amounts of mechanically comminuted material. Each of these curves may be regarded as a composite formed by the combination of a curve of the type of V with a curve of the type shown in *Fig. 1*. In Curve I, the mechanically disintegrated material predominates, whilst in Curve VI, there is only a slight indication of such material in the upper part of the range.

Curve II is of interest because it relates to a comparatively juvenile soil in which comminution has been mainly mechanical. It resembles the curves of *Fig. 1*, but has a greater range, and shows a trend similar to that of the other soil curves at the lower end.

SIGNIFICANCE OF MECHANICAL ANALYSIS AT LOWER END OF PARTICLE RANGE.

We must now refer to a consideration which affects the significance of the data for the finer fractions in mechanical analysis, namely, the effect of the cation associated with the dispersed particles. Dispersion is usually effected in an alkaline medium such as dilute ammonia or dilute sodium

hydroxide solution. According to the cation used, we may refer to an ammonium soil, a sodium soil, or, if the soil has been bereft of associated bases by acid treatment and prolonged washing, a hydrogen soil. Inasmuch as ions vary in their degree of hydration, the particles of a suspension subjected to mechanical analysis will be invested with aqueous shells of varying thickness. It has been shown by G. W. Robinson⁶ that, if v be the settling velocity of a particle of radius r , the effect of an aqueous sheath of thickness d will be to reduce the velocity to $vr/(r+d)$. For particles down to the limits of an ordinary mechanical analysis, the aqueous sheath effect is scarcely appreciable, and sodium, ammonium, and hydrogen suspensions yield practically the same results. With particles of smaller size, however, differences in cation hydration lead to different results being obtained with different cations. Comparing a heavily hydrated cation such as sodium with a slightly hydrated cation such as hydrogen, it is evident that a particle associated with sodium ions will carry a larger aqueous sheath than a particle associated with hydrogen ions, and, if the two particles are of equal size, the sodium particle will have a smaller settling velocity. (See p. 113 *et seq.*)

Below the ordinary clay limit, therefore, the mechanical composition for hydrogen soil will give a truer picture than that for sodium soil, in which the aqueous sheath effect tends to make particles settle more slowly and therefore, assuming the ordinary size velocity relationship, appear smaller than their actual dimensions. C. E. Marshall⁷ has suggested that clays associated with different cations may actually have different degrees of dispersion and that a particle of sodium clay is actually smaller than a particle of, say, hydrogen clay.

THE CLAY FRACTION.

If the fractions distinguished in mechanical analysis be isolated by using a decantation method it will be seen that they exhibit important differences. Whilst the coarser fractions, namely, the sands and the silt, consist usually and

mainly of original fragments such as quartz, feldspars, and micas, the clay fraction consists almost entirely of material of secondary origin and of colloidal character. It should be noted, however, that in materials comminuted by mechanical means, although appreciable proportions of the clay fraction occur, it is not in such cases colloidal in character. For example, the slate dust represented by Curve V in *Fig. 1* contains 59% of clay fraction, but is lacking in the typical properties of colloidal clay.

The colloidal properties which distinguish the clay fraction of soils most markedly from sand and silt may be briefly enumerated. Firstly, colloidal clay exhibits in a high degree the property of retaining water by imbibition. We shall consider this more fully in discussing the water relationships of soils, but it may be noted that the imbibition of water by clay is essentially similar to that exhibited by such typical colloidal gels as gelatin, agar, and silicic acid. The presence of considerable proportions of clay confers a high water-holding capacity on soils and it is to be noted that, whilst the water retained by non-colloidal sand and silt is in the form of surface films, the water associated with clay is held within the cellules of the colloid as vesicular water.

Secondly, changes in the water content of colloidal clay are accompanied by marked changes in volume. This is plainly demonstrated by the cracks which develop on drying. Indeed, the shrinkage of highly colloidal soils on drying out may lead to the development of a marked columnar structure whereby the soil breaks up into prisms of polygonal cross-section.

Thirdly, colloidal clay exhibits the property of plasticity when associated with certain proportions of water. In this respect it is sharply distinguished from sand and silt, for even the finest silt is devoid of this property.

Fourthly, colloidal clay confers cohesive properties on soils. If a quantity of silt free from clay be moistened and moulded into the form of a brick, it will be found that the

brick crumbles under the slightest stress on drying. Where colloidal clay is present, even in comparatively small proportions, the dried moulded soil exhibits considerable resistance to applied stresses. This is illustrated by the difficulties attending the tillage of clay soils which, if allowed to dry into clods, are no longer amenable to the ordinary operations of tillage.

Colloidal clay occurs associated with colloidal organic matter and is present in the soil partly as independent aggregates and partly as coatings investing the non-colloidal particles. In this way it renders possible the existence of compound-particle structure and thereby confers on soils the capacity for *flocculation* and *deflocculation*. Colloidal clay is highly sensitive to the presence of small quantities of electrolytes. In the presence of neutral salts it tends to form compound aggregates and is said to flocculate. On the other hand, in the presence of weak alkaline solutions, it tends to disperse into a state in which the ultimate particles behave independently and in which its colloidal properties are most strongly manifested. Certain soils which contain sodium carbonate are highly impervious to water, and are sticky and intractable under tillage operations, which may even become impossible. In such soils, the individual particles exist independently and there is little or no aggregation into compound particles or crumbs.

Finally, the colloidal clay is pre-eminently the reactive constituent of the mineral portion of the soil. It has the typical colloidal property of absorbing certain substances from solution. The absorptive properties of soils, however, reside both in the colloidal clay and in the colloidal organic matter, which are, for most purposes, conveniently considered together as a single *colloidal complex*. The reactions of this complex will be considered in greater detail in a later chapter.

Whilst the non-colloidal particles of the soil are almost entirely crystalline fragments of rock forming minerals, the colloidal clay has been generally considered to be amorphous.

Recent studies, however, by the method of X-ray analysis have shown that in many clays the colloidal material is actually crystalline and built up of small platelets. Such a condition is not inconsistent with gel character, for many colloidal gels have already been shown to consist of crystalline elements.

In this connexion it is noteworthy that R. Bradfield⁸ and others have shown that, for the same soil, the colloidal clay is apparently of uniform composition. A. F. Joseph⁹ has shown that the fractionation of colloidal clay by the use of different settling velocities does not actually reveal distinct fractions, but that the coarsest particles within the clay fraction can be further dispersed. It may be, therefore, that the conception of a range of particles of varying size obtained from a consideration of the non-colloidal fractions is not transferable to the clay fraction and that the coarser grades in this fraction are actually compound particles built up of constituents of similar lattice structure.

ORGANIC MATTER.

Whilst soil exhibits, so far as its mineral portion is concerned, a close resemblance to other unconsolidated products of superficial weathering and even to non-metamorphosed sediments which are classed geologically as rocks, the presence of organic matter confers on the soil its distinctive character as the seat of plant life. It is true that some sedimentary rocks contain organic matter; but the organic matter in such cases, though doubtless originally similar to that in the soil, has become considerably modified by secular changes. Whilst the soil organic matter is the habitat of a complex micro-flora and fauna, the organic matter of such sediments as bituminous shales and coal is completely devoid of microbiological activity and, if brought under the influence of soil micro-organisms, is as indifferent to their activities as so much inert mineral matter.

The altered organic matter of such sediments as shales and coals also differs from soil organic matter in the much

higher proportion of carbon which it contains. Whilst soil organic matter contains from 55 to 60% of carbon, the organic matter of shales contains much higher proportions; and in the extreme cases of anthracites, is almost entirely carbon.

The proportion of organic matter in soil varies considerably. Whilst in certain tropical soils it may fall below 1%, it may constitute practically the whole of the dry matter of peats. The factors influencing the amount of organic matter in soils will be discussed more fully in a later chapter.

The determination of the exact proportion of organic matter in soil is fraught with considerable difficulty. The loss on ignition is frequently reported as a measure of organic matter; but the figure is almost valueless in the majority of cases since it includes chemically combined water in the mineral portion of the soil and, in calcareous soils, the carbon dioxide driven off from carbonates. And thus it may happen that a heavy soil practically devoid of organic matter may show a loss on ignition of 10 to 12%, consisting almost entirely of chemically combined water driven off at the temperature of ignition. Highly calcareous soils may also show ignition losses out of proportion to the amount of organic matter which they contain. The error in using loss on ignition as a measure of the organic matter is least serious in the case of sandy soils; but, even here, the determination of small quantities is subject to considerable error.

A more satisfactory method of estimating the amount of organic matter in a soil is by determining the amount of organic carbon present. This can be carried out either by direct combustion or by one of the methods of wet oxidation. Where the amount of carbon dioxide evolved is measured, it is necessary to apply a correction in the case of calcareous soils, preferably by removing the calcium carbonate in a preliminary treatment. To obtain the amount of organic matter, the figure for organic carbon is multiplied by a conventional factor, usually 1.724, obtained on the assumption that soil organic matter contains 58% of carbon. The use

of this factor is open to considerable criticism, for it is based on the composition of the organic matter of peaty soils and not on the composition of the organic matter of ordinary soils. Since no method is known or is likely to be discovered whereby the organic matter of soils can be isolated without alteration, it seems clear that some doubt must always exist as to the exact factor to be used. For comparative purposes, however, the use of the organic carbon itself as a measure of organic matter would appear to be the most useful procedure.

W. O. Robinson¹⁰ has proposed the estimation of soil organic matter by treating the soil with 15-30% hydrogen peroxide and determining the loss. The objection to this method is that the oxidation is not always completely quantitative, and also that it cannot be assumed that the treatment is without effect on the mineral portion of the soil. The present writer has found that, even with 6% hydrogen peroxide, the oxidation of organic matter is accompanied by a certain solution of mineral matter, principally sesquioxides, consequent on the formation of organic acids. Pyridine and acetyl bromide have been proposed as solvents for organic matter, but they have not as yet come into general use.

The organic matter of the soil consists partly of recognisable fragments of plant material, and partly of material which has lost all traces of its original structure and has become changed into dark amorphous humified matter or humus. Whilst in certain cases, such as the superficial layers of grass soils, the undecomposed material with recognisable structure may form a considerable proportion of the total organic matter, usually 80% or more consists of dark coloured structureless material which has been derived from added plant residues by a complex series of microbiological and, to some extent, chemical decompositions.

A certain indefiniteness exists in the terminology applied to soil organic matter. Whilst many writers use humus as synonymous with organic matter, others have applied the

term to a definite fraction, for example, the so-called *matière noire* of Grandeau, soluble in 4% ammonia. This confusion has led to the abandonment of the term humus by many writers. In the opinion of the present writer the name humus might be conveniently retained for the organic matter of the soil which has been decomposed and has lost its original structure. Humus would thus exclude recognisable fragments of plant materials which have not yet become part of the characteristic organic matter of the soil. The fact that there is no generally received method for determining humus, as distinct from total organic matter, should not prevent us from recognising its existence and referring to it by an appropriate name.

Much of what has been said of the colloidal clay applies to humus, which is also of a colloidal character. Its characteristic properties are exhibited in an eminent degree by peat, which is distinguished by its high capacity for absorbing water and also shows great changes in volume on wetting and drying.

Plasticity does not appear to be a distinctive character of colloidal organic matter, whilst cohesion is shown to a much smaller degree than by clay. Indeed, the presence of sufficient humus in the soil may partially neutralise some of the undesirable physical consequences of a high proportion of colloidal clay. At the same time, humic matter confers sufficient cohesion to render possible the formation of compound aggregates or soil crumbs in sandy soils and to that extent exerts a favourable effect.

Humus is requisite for the microbiological processes of the soil and is, with its associated moisture, the medium in which soil organisms fulfil their life cycles. The presence of humus and its living population is the distinctive character of a soil, for, whilst raw subsoils contain colloidal clay, and even small amounts of organic matter, they are, below moderate depths, devoid of living organisms.

The great complexity of soil organic matter has long been realised. In a later chapter we shall discuss its con-

stitution in connexion with the decomposition of plant residues in the soil. We may note, however, that the greater part of the soil humus consists of relatively inert material whose rôle in the soil is mainly physical. The material directly concerned in microbiological changes includes undecomposed plant residues, and a comparatively small fraction of the humus, consisting of simple and presumably active compounds.

THE COLLOIDAL COMPLEX.

We have, up to the present, for convenience of presentation, considered the colloidal clay and the humus separately. From the point of view of the constitution of the soil, however, they cannot be considered apart, since they are associated together, if not in actual chemical combination, in the most intimate physical relationship, and form the *colloidal complex* of the soil. Whilst the non-clay mineral matter of the soil exhibits in some degree those properties of colloids associated with surface development, it is in the clay-humus complex that these properties are principally seated.

Since colloidal properties such as imbibition of water, swelling, and absorption, are much more pronounced in organic matter than in clay, it is obvious that the gross amount of colloidal matter present is no true measure of the colloidity of the soil. A soil containing 25% of colloidal material, made up of 20% of clay and 5% of organic matter, will differ markedly from a soil also containing 25% of colloidal matter, but made up of 5% of clay and 20% of organic matter. Data purporting to give the percentage of "colloid" in the soil should, therefore, be interpreted with caution.

The colloidal constituent of soils occurs usually in films or coatings investing the non-colloidal particles. In highly colloidal soils, however, it is probably present also in the form of independent aggregates in which, doubtless, fragments of crystalloidal material may be embedded. It is

unusual to find soils entirely composed of colloidal matter, but certain highly weathered tropical soils may contain as much as 90%, of which nearly the whole is colloidal clay. The non-colloidal material in such soils consists of unweatherable minerals such as quartz and magnetite. In extreme cases, peat soils may consist almost entirely of organic matter, of which the greater proportion is of colloidal character.

It is worthy of note that when soils are prepared for laboratory examination by light grinding and sieving, the larger aggregates in the sample contain more colloidal clay and organic matter than the finer material. For example, A. M. Smith¹¹ found that aggregates larger than 1 mm. in diameter in a 3 mm. sample of soil had a greater content of exchangeable bases than the finer material.

SOIL SOLUTION.

The study of the soil solution offers special difficulties from the fact that the soil cannot be regarded simply as a mixture of insoluble inert material with substances of varying solubility. For example, in the case of a mixture of quartz sand and calcium sulphate, the concentration of the liquid phase in equilibrium would remain constant as long as any calcium sulphate remained undissolved. Thereafter, the concentration would vary inversely as the volume of the liquid phase. The situation in the soil is different because of the reactivity of the insoluble solid phase, which exhibits the phenomena of absorption and base exchange.

At the dilutions ordinarily encountered in the soil solution, the salts present are almost entirely dissociated into their component ions. In considering the equilibria between the solid phase of the soil and the soil solution, it is therefore necessary to think in terms of ions rather than salts. Thus, the soil solution has a certain concentration of calcium-ion and this concentration will vary with the ratio of soil to water. The soil calcium with which the calcium of the soil solution is in equilibrium is the exchangeable

calcium, combined with the insoluble acids of the clay and organic matter.

If y be the concentration of calcium in the soil phase, and C the concentration in the soil solution phase, it is found that $y = K.C^{1/p}$, K and p being constants. This is the well known Freundlich equation, which applies to the adsorption of substances from solutions by colloids. The equilibria of most ions are governed by the same law, although with different constants. The fact that experimental results fit the Freundlich equation must not, however, be taken as proving that these equilibria are colloidal adsorptions analogous to the bodily adsorption of dyes by charcoal, for the equation will serve equally well for many non-colloidal heterogeneous equilibria.

Certain ions do not obey the Freundlich equation, notably chloride, and nitrate. These ions are not ordinarily absorbed by soils and there is, thus, no soil chloride, or soil nitrate with which they are in equilibrium. The concentration of these ions in the soil solution, in the absence of any removal by leaching or root absorption, will therefore be inversely proportional to the ratio of water to soil.

The above considerations should make it clear why the soil solution cannot be investigated simply by adding a known volume of water to a given weight of soil, filtering, and analysing an aliquot of the filtrate for the purpose of calculating back to the original water content of the soil.

Various methods have been proposed for isolating the actual soil solution. The application of great pressure has been used for the purpose of expressing a portion of the soil solution for analysis, but has not proved a convenient method in practice. Another method which has been used for obtaining the soil solution is the use of centrifugal force. Moist soil is centrifuged at high speed and a certain proportion of the solution is thrown out. In this method, as in the pressure method, a considerable proportion of the soil solution may be left in the soil if it is highly colloidal.

The most convenient methods for obtaining the soil solution depend on displacement by other liquids. Moist soil is packed into suitable cylinders with perforated bottoms and the surface of the soil is flooded with a suitable liquid. This percolates into the soil, driving the soil solution before it. Eventually the lower part of the soil column becomes sufficiently saturated with the displaced soil solution for an excess to drain away. Drainage and collection are allowed to continue until the displacing liquid itself begins to percolate through. In the original experiments of Schloesing in 1866, water coloured with carmine was used, the appearance of the dye in the percolate indicating the point at which the original soil solution had been displaced. Paraffin oil and ethyl alcohol have also been used.

The results expressed as parts per million, obtained by Schloesing for a soil containing about 20% of moisture, are as follows:—

SiO ₂	N ₂ O ₅	CO ₂	CaO	MgO	K ₂ O	Na ₂ O	SO ₃	Cl	Organic Matter
29.1	305	118	264	13.5	6.9	7.8	57.9	7.4	37.5

J. S. Burd and J. C. Martin¹² have examined the soil solution of certain Californian soils, using a method of displacement under pressure. Some results for a soil at the beginning and end of the crop season are given in Table II.

TABLE II. — COMPOSITION OF SOIL SOLUTION (BURD AND MARTIN).

Date	pH	Parts per million in displaced solution								Total Solids
		NO ₃	HCO ₃	SO ₄	PO ₄	Ca	Mg	Na	K	
April 30th, 1923	7.4	149	83	561	1.1	242	91	42	21	1190
Sept. 4th, 1923	7.6	58	155	432	0.6	193	47	40	9	935

The results are of the same order of magnitude as those obtained by Schloesing, the most notable difference being the higher sodium content in the solutions from the

Californian soil. The decrease in concentration consequent on the growth of a crop may be noted.

The figures given merely serve to indicate the order of magnitude of the concentration of the principal ions. It may be expected that the concentration of the soil solution will increase with decreasing moisture content, and will be greater in dry climates than in humid climates, where soluble materials are leached out by percolating waters and lost in drainage.

There are undoubted difficulties in the way of displacement methods, which in practice prove to be of limited application. The moist soil must be well packed in the displacement cylinder in order to avoid percolation of the displacing liquid through wide capillaries. This is an obstacle in the way of investigation of heavy soils since any attempt at packing results in complete hindrance of percolation.

TABLE III.—ANALYSES OF DRAINAGE WATER FROM CULTIVATED FIELDS, PARTS PER MILLION (A. VOELCKER).

Constituent			Treatment		
			Unmanured	Farmyard manure	Complete artificial fertilisers
CaO	...	...	98.1	147.4	143.9
MgO	...	...	5.1	4.9	7.9
K ₂ O	...	...	1.7	5.4	4.4
Na ₂ O	...	...	6.0	13.7	10.7
Fe ₂ O ₃	...	...	5.7	2.6	2.7
Cl	...	...	10.7	20.7	20.7
SO ₃	...	...	24.7	106.1	73.3
P ₂ O ₅	...	...	0.6	...	1.54
SiO ₂	...	...	10.9	35.7	24.7
N as NH ₃	...	...	0.14	0.20	0.24
N as Nitrate	...	...	15.0	62.0	32.9
Org. Matter, CO ₂	...	...	...	...	...
etc.	...	...	67.7	77.3	84.6
Total Solids	...	...	246.4	476.0	407.6

Drainage waters, though more dilute than the soil solution, show a general resemblance in composition. Abundant data are available, but perhaps the most instructive are those obtained from the Broadbalk wheat field at Rothamsted, where drainage is collected from plots which have been continuously under wheat for about 80 years. Table III gives the data for drainage from three of the plots obtained by A. Voelcker¹³. The results are expressed in parts per million.

SOIL AIR.

Like the soil solution, the soil air is a dependent phase and its composition varies even in the same soil. The chief respects in which it differs from the free atmosphere are: (1) it contains a much larger proportion of carbon dioxide; (2) it is, except in air-dry soils, saturated with water vapour; and (3) it contains rather less oxygen and nitrogen.

The highest proportions of carbon dioxide are found in soils rich in organic matter. For example, the air of grassland soils may contain as much as 1.5% of this constituent. Arable soils may contain up to 0.5%, the highest figures being found in soil recently treated with organic manures.

In the same soil, the proportion of carbon dioxide in the soil air will vary with the energy of microbiological decomposition. In Britain, where the soil generally contains sufficient moisture for micro-organisms to fulfil their functions, microbiological activity in a given soil is controlled mainly by temperature and we should therefore expect the greatest concentration of carbon dioxide to be found during the summer months. Aeration also plays a part in the rate of production of carbon dioxide, since the oxidative decomposition of organic matter is an aerobic process. The amount of carbon dioxide in the soil will not, of course, depend simply on the rate at which it is produced by microbiological activity, for diffusion will also enter as a factor.*

*For an investigation of soil air under English conditions see E. J. Russell and A. Appleyard, *J. Agric. Sci.*, 1915, 7, pp. 1-48.

It has been suggested that the rate of evolution of carbon dioxide from the soil is a measure of biological activity and hence of fertility. Whilst this may be true in comparing soils of similar constitution under similar conditions, it is not likely that the rate of production of carbon dioxide can serve as a general measure of fertility.

The carbon dioxide produced in the soil from the biological decomposition of organic matter and from the respiration of plant roots constitutes the principal agency whereby the supply of this constituent, depleted by photosynthesis, is regenerated in the atmosphere. It is also of importance as a constituent of soil moisture and of percolating waters, and thus contributes to the formation of soil by intensifying the hydrolytic action of water on mineral silicates.

SOIL STRUCTURE.

We have seen that the principal constituents of the soil are (1) mineral matter, consisting of particles varying in size from stones to clay; (2) organic matter, consisting of decayed and decaying residues of vegetation and associated with clay to form the colloidal complex; (3) soil solution, which is a dilute solution of the bicarbonates, sulphates, chlorides, nitrates, phosphates and silicates of calcium, magnesium, potassium, sodium and iron; and (4) soil air, which differs from atmospheric air in being saturated with water vapour and enriched in carbon dioxide. It now remains to consider the structure of the soil.

Soil being a granular substance, a given volume of soil will include a certain proportion of interstitial spaces or *pore space*, which may be filled either with water or air. We may therefore distinguish the apparent volume of a certain quantity of soil from the true volume occupied by the particles which compose it.

If 100 g. of soil occupy a volume, V c.c., then the apparent density, d_1 , is $100/V$. Now, if the true density of the soil be d , its true volume is $100/d$ and the volume occu-

pied by the interstitial space is $V - 100/d$ or $100/d_1 - 100/d$, which, expressed as a percentage, is $100(d - d_1)/d$.

If a soil could be imagined to consist of uniform spherical particles with closest packing, the pore space would be 26%. In practice this condition is never fulfilled, for not only are there particles of different sizes present, but, owing to the presence of colloidal material, the ultimate particles are to a large extent aggregated together to form compound particles or soil crumbs, so that the pore space is actually never so low as in the case of uniform spheres with close packing. It generally varies from 40 to 60%.

The pore space is not a specific property of a given soil, but depends on its temporary structure. Cultivation has the effect of increasing pore space by encouraging the formation of compound aggregates, whilst, on the other hand, pore space may be diminished by pressure or treading. In ordinary cultivated soils the pore space is at maximum after ploughing and gradually diminishes as the soil settles down under the influence of weather. The effect of frost is to increase pore space.

The proportion of pore space in a soil depends on the extent to which its primary particles are united together into compound particles or crumbs. This possibility is greatest in soils containing large proportions of clay or organic matter and least in barren sands and gravels. Inasmuch as a high pore space is favourable to aeration and water movements, it is obvious that the crumb structure is most desirable for plant growth. On the other hand, the destruction of crumb structure and the prevalence of the so-called single grain structure leads to a diminution of pore space and unfavourable conditions for plant growth.

Structure is an important characteristic in the study of the soil profile. The Russian pedologists, who have given the greatest attention to this phase of the subject, distinguish a large number of different structure forms, such as single-grain, granular, prismatic, columnar, nuclear, and structureless. Profile descriptions should always include observa-

tions on structure, for many types of soil are characterised by this property. The top horizons of podsoils generally show a single grain structure. Black earths show a porous granular structure in the surface soil and columnar or prismatic structure in the lower horizons. Alkaline soils show a columnar or prismatic structure, whilst saline soils are described as structureless. The surface horizon of laterites often exhibits a vesicular or honeycomb structure.

A. T. Tiulin¹⁴ distinguishes two types of compound aggregates in soils, namely, those in which the cementing or flocculating agents are bivalent and trivalent cations, and aggregates held together by univalent cations or by simple cohesion. The former are termed *true aggregates* and the latter *false aggregates*. The preliminary dispersive treatment in mechanical analysis is required to break down true aggregates. It is suggested by Tiulin that a measure of the proportion of true aggregates may be obtained by sieving a soil in water, whilst the proportion of false aggregates may be obtained by dry sieving.

BIBLIOGRAPHY.

- ¹ KRAUSS, G., "Ergänzender Bericht über eine dem Prager bodenkund. Kongress vorgetragene neue Methode der mechanische Bodenanalyse usw." *Int. Mitt. Bodenk.*, 1923, **13**, pp. 147-160.
- JENNINGS, D. S., THOMAS, M. D., and GARDNER, W., "A new method of mechanical analysis of soils." *Soil Sci.*, 1922, **14**, pp. 485-499.
- ROBINSON, G. W., "A new method for the mechanical analysis of soils and other dispersions." *J. Agric. Sci.*, 1922, **12**, pp. 287-291.
- ² ODÉN, S., "Eine neue Methode zur mechanische Bodenanalyse," *Int. Mitt. Bodenk.*, 1915, **5**, pp. 257-311.
- ³ WIEGNER, G., "Über eine neue Methode der Schlammanalyse," *Landw. Vers. Stat.*, 1918, **91**, pp. 41-79.
- ⁴ ROBINSON, G. W., "On the form of mechanical composition curves of soils and other granular substances." *J. Agric. Sci.*, 1924, **14**, pp. 626-633.
- ⁵ SHAW, W., "Quarry products for modern roads." *Quarry Managers' Journal*, 1925, **8**, p. 57.
- ⁶ ROBINSON, G. W., "Note on the mechanical analysis of humus soils." *J. Agric. Sci.*, 1922, **12**, pp. 396-321.
- ⁷ MARSHALL, C. E., "The orientation of anisotropic particles in an electrical field." Part I.—General. Part II.—The application to determination of the double refraction of clays. *Trans. Faraday Soc.*, 1930, **26**, pp. 173-189.

- ⁸ BRADFELD, R., "The chemical nature of colloidal clay." *J. Amer. Soc. Agronomy*, 1925, **17**, pp. 253-270.
- ⁹ JOSEPH, A. F., "Clays as soil colloids." *Soil Sci.*, 1925, **20**, pp. 89-94.
- ¹⁰ ROBINSON, W. O., "The determination of organic matter in soil by means of hydrogen peroxide." *J. Agric. Res.*, 1927, **34**, pp. 339-356.
- ¹¹ SMITH, A. M., "Exchangeable bases in some Scottish soils." *J. Agric. Sci.*, 1928, **18**, pp. 67-75.
- ¹² BURD, J. S., and MARTIN, J. C., "Secular and seasonal changes in the soil solution." *Soil Sci.*, 1924, **18**, pp. 151-167.
- ¹³ VOELCKER, A., "On the productive power of soils in relation to the loss of plant food by drainage." *Trans. Chem. Soc.*, 1871, **24**, pp. 276-297.
- ¹⁴ TIULIN, A. T., "Aggregate analysis as a method for determination of real soil structure." *Perm. Agric. Exp. Stat. Div. Agric. Chem. Report*, 1928, pp. 1-24, 77-122.

CHAPTER III.

THE PEDOGENIC PROCESSES.

SCOPE OF ENQUIRY.

IN the present chapter, attention will be devoted mainly to the processes whereby the mineral soil profile is developed. The addition of organic matter and its transformations within the soil are, of course, of great importance, both on account of the contribution of humus to the characteristic properties of soils, and also on account of the part played by humus in determining the character of the leaching processes, which affect the type of profile development. But, after considerable deliberation, it was felt by the writer that the subject of soil organic matter could be more conveniently handled in a separate chapter. In the present chapter, therefore, organic matter is only discussed in so far as it is necessary for an adequate comprehension of the transformations affecting the mineral matter of the soil.

The interest of the student of soils in the processes of rock decay does not extend over the whole field of the subject. He is not directly concerned with those processes occurring in what has been described by C. R. Van Hise¹ as the zone of anamorphism. Such processes as the production of kaolin by deep-seated hydrothermal action are not primarily of importance for the study of soils. The weathering processes which may be fitly described as pedogenic or soil-forming are those which are directly due to the action of atmospheric agencies on superficial materials.

PHYSICAL WEATHERING.

Two types of weathering may be distinguished, namely, (a) *physical or mechanical weathering*, and (b) *chemical weathering*. The processes whereby rocks are physically weathered are familiar to students of geology. They are essentially processes which result merely in a comminution of the materials on which they operate. Granite which has simply been subjected to mechanical weathering is still granite, except in so far as the space mosaic of constituent minerals has been broken up. There is no question of chemical change or the formation of new products.

The most universal type of physical weathering is that produced by changes in temperature. Where such changes are of such sufficient magnitude and suddenness, expansion or contraction in superficial layers of rocks can result in strains which ultimately lead to shattering. This type of weathering is most marked in dry climates, in which free radiation results in great changes in temperature at sunrise and sunset. The weathering of rocks in deserts is predominantly of this type.

In cool or temperate climates with abundant precipitation, the alternations of frost and thaw are of chief importance in the physical comminution of rocks. This action depends on the occurrence in rocks of joints and interstices which, during winter, become filled with water. The screes of our mountain valleys are striking evidences of the disruptive action of alternating frost and thaw, and the same processes are continued in every fragment which, by its porosity, is capable of holding water.

Other agencies of physical weathering are moving water in rivers, moving ice in glaciers, wave action, and the blast of sand grains transported by wind. For a detailed study of these processes the student may be referred to any textbook of geology, to which their study more particularly belongs.

Soils in the early stages of development, desert soils, and the soils of the arctic and alpine regions are pre-

dominantly the result of physical weathering. They are generally characterised by the presence of large proportions of coarse material, and their mechanical composition curves often show an approximation to a linear form in their upper range.

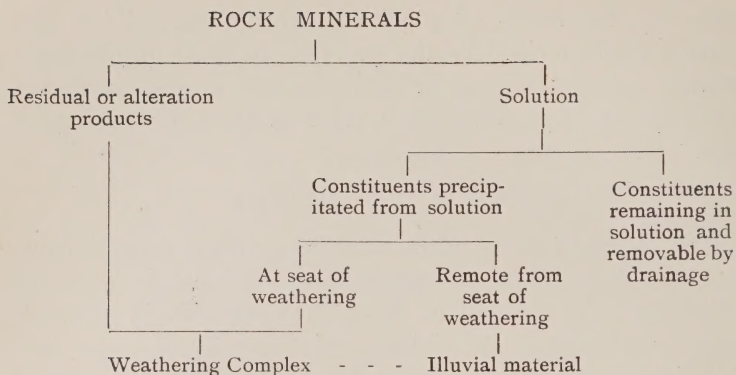
CHEMICAL WEATHERING.

The chemical weathering of rocks involves more serious alterations than the mere comminution produced by physical weathering. As a result of these changes, certain minerals disappear wholly or partially, and material of secondary origin is formed which differs markedly from the parent minerals.

The processes embraced under this head fulfil themselves in an aqueous medium and depend on the decomposing action of water, reinforced by the presence of dissolved carbon dioxide and, in some cases, organic acids formed from the decay of plant residues. Since chemical weathering takes place at the surface of rock minerals, it is evident that it is intensified where physical comminution has preceded it. But since most rocks are formed of an irregular space mosaic of different minerals with different degrees of susceptibility to attack, chemical weathering alone is capable of producing disintegration, particularly where there is a certain degree of jointing or porosity in the weathering rock.

Essentially, chemical weathering involves two phases, namely, (a) the disappearance of certain minerals, and (b) the formation of secondary products. Some of the secondary products may originate by alteration *in situ* of the parent minerals, whilst other products may originate by precipitation from the soluble products of weathering. Such precipitation may occur at the seat of weathering, or remote from the seat of weathering after transport by percolating water. Material precipitated at the seat of weathering may be mixed, or even enter into combination, with residual products.

The processes may be represented diagrammatically as follows :—



The *weathering complex*, representing that portion of the soil which is of secondary origin, may include residual or alteration products and material precipitated *in situ* from soluble products of chemical weathering. It may also include material transported from elsewhere and reprecipitated. (See further, Chapter IV.)

The chemical weathering of rock affects principally the mineral silicates, such as feldspars, micas, and ferromagnesian minerals, and depends on their instability at ordinary temperatures in the presence of water and carbon dioxide. From data by F. W. Clarke², it is estimated that the approximate proportions of the principal minerals in the earth's crust are as follows:—Feldspars 57.8%; Amphiboles and Pyroxenes 16.0%; Quartz 12.7%; Micas 3.6%. It is thus evident that the feldspars are by far the most important group of soil-forming minerals, followed by the ferromagnesian minerals. Certain minerals are practically unaffected by chemical weathering and persist unaltered in the soil. Of these the principal example is quartz. Other minerals resistant to chemical weathering are magnetite, titanite, ilmenite, etc.

The chemical weathering of rocks depends on the presence of water. Since the rate of chemical change is affected by temperature, it will be recessive at low temperatures and most rapid at high temperatures. Translated into terms of

natural conditions, this means that chemical weathering is at minimum under desert conditions, owing to absence of water, and under arctic and alpine conditions, owing to the prevalence of low temperatures. In such cases the weathering which occurs is almost entirely of a physical character. In humid climates, the intensity of weathering will increase with temperature and reach a maximum in the tropics.

Soils predominantly of the physically weathered type may occur in a region favourable for chemical weathering. Such soils are generally young soils derived from crystalline rocks and are developed on recent deposits or in situations where erosion is sufficiently intense to prevent the accumulation of the products of decay and the consequent predominance of chemical over physical weathering.

SIMPLE SOLUTION.

Simple solution involving the total removal of the weathering material is negligible except in the case of calcium carbonate which, compared with silicate minerals, is very soluble in percolating waters, particularly when such waters contain dissolved carbon dioxide which has originated from the respiration of plant roots and the decomposition of organic residues in the soil.

The more spectacular effects of the solution of calcium carbonate are seen in the cave systems which are so characteristic of limestone countries. From the point of view of soil formation, the solution of calcium carbonate is of interest by reason of the fact that limestones generally contain an admixture of non-calcareous material which remains behind as a residue. Limestone soils are thus essentially residual soils, and their character depends principally on the nature of the material associated with the calcium carbonate in the parent limestone. Thus we may have soils of all textures from sands to clays originating from the weathering of calcareous rocks.

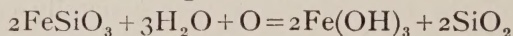
Soil formation from limestones cannot always be explained as a simple removal of calcium carbonate. During

the process of liberation of the residual material certain changes may occur which profoundly modify its character. The terra rossa soils associated with limestones in the Mediterranean region represent material which has been considerably modified from the parent residue.

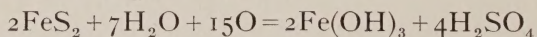
HYDRATION AND OXIDATION.

Although the silicate minerals are considerably less soluble than calcium carbonate, yet solution, or, at least, the presence of water, is a necessary circumstance in their chemical decomposition. Before dealing with the most important type of chemical weathering of silicates consequent on solution, we must notice two processes in which, although water is necessary for their fulfilment, actual solution is not demonstrably required. *Hydration* consists in the chemical combination of water with a particular mineral. An example of this is the formation of serpentine from olivine. In itself hydration is not essentially a pedogenic process since it occurs also under conditions remote from the operation of ordinary atmospheric agencies. It may, however, be of importance as a preliminary to the more characteristic pedogenic processes.

Oxidative processes, broadly speaking, only affect ferrous compounds. Certain ferromagnesian compounds contain ferrous iron, which, in the presence of air and moisture, is capable of being oxidised with the production of hydrated ferric oxide, e.g.,



Iron pyrites and marcasite yield by oxidation hydrated ferric oxide and sulphuric acid :—



HYDROLYSIS.

We may now consider the most important of the processes whereby minerals are weathered chemically, namely, hydrolysis. Hydrolysis is a consequence of the partial dis-

sociation of water into hydrogen-ions and hydroxyl-ions. The actual amount of dissociation is exceedingly small, for at ordinary temperatures a litre of water contains only 10^{-7} g. of hydrogen in the ionic form; but the presence of carbon dioxide in solution increases the hydrogen-ion concentration, and thus reinforces the hydrolytic action of the water.

The presence of hydrogen-ions is characteristic of acid solutions. Strong acids such as hydrochloric and sulphuric acids in dilute solutions are almost completely dissociated. Thus, in a decinormal solution of hydrochloric acid, practically all the hydrogen is in the ionic form, and the hydrogen-ion concentration is nearly 0.1 g. per litre. In a weak acid, on the other hand, dissociation is only partial. A decinormal solution of acetic acid is only slightly dissociated and the hydrogen-ion concentration is about 0.0013 g. per litre.

Water thus behaves as a weak acid, and its effect on weatherable silicate minerals depends on the activity of the hydrogen-ions. The minerals with which we are concerned may be regarded as complex silicates of calcium, magnesium, potassium, sodium, aluminium, and iron. The simplest type of decomposition would consist in the replacement of alkali or alkaline earth-ions in the mineral lattice by hydrogen-ions, resulting in the formation of aluminosilicic or ferrosilicic acids and the liberation of alkali or alkaline earth hydroxides. A further stage in breakdown might consist in the separation of silicic acid, accompanied by a profound modification of the original crystal lattice. A process such as the following may be conceived to occur in the formation of kaolinite from potash felspar :—

(1) $K_2O \cdot Al_2O_3 \cdot 6SiO_2 + 2H_2O = H_2O \cdot Al_2O_3 \cdot 6SiO_2 + 2KOH$
followed by :—

(2) $H_2O \cdot Al_2O_3 \cdot 6SiO_2 + H_2O = Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O + 4SiO_2$

The action of water on felspar is illustrated by the classical experiment of Daubrée, in which 3 kg. of orthoclase felspar, after 200 hours' agitation with water, yielded 2.52 g. of K_2O in solution and an insoluble alteration product probably of a kaolin-like character. Under laboratory con-

ditions, however, the hydrolysis of silicates, as evidenced by the development of an alkaline reaction, slows down after its initial stages owing to the accumulation of a protecting layer of the insoluble products of the reaction.

The hydrolysis of mineral silicates is more commonly supposed to take place entirely in solution. Although the original minerals are highly insoluble, the consequences of such slight solubility as they possess may be highly significant over long periods of time.

It is supposed that the first stage consists in the simultaneous liberation of silicic acid and the hydroxides of the bases present. Whilst the alkaline and alkaline earth hydroxides are soluble in water, the silicic acid and the hydroxides of aluminium and iron may form colloidal solutions. Silicic acid sols are electronegative, whilst the sesquioxide sols are electropositive, and mutual precipitation may occur, giving rise to the weathering complex. This may take place in one of the following ways:—

(a). The mutual precipitation may, as originally suggested by J. M. Van Bemmelen³, result in the formation of amorphous *absorption compounds* of indefinite composition. The general formula for such compounds might be written,
 a SiO₂. b Al₂O₃. c Fe₂O₃. d CaO. e MgO. f K₂O. g Na₂O.
 h H₂O

(b). The absorption compounds of indefinite composition and amorphous structure initially formed may give rise to crystalline compounds of definite, though complex, composition.

(c). The mutual precipitation may give rise directly to crystalline compounds of definite, though complex, composition.

Recent work by S. Mattson, to be discussed in a later stage of this book, has shown that substances resembling the weathering complexes of soils may be formed by mutual precipitation of colloidal solutions of opposite electrical sign and that their composition can be influenced by the reaction of the medium and the nature of the cations present.

The ingenuity of Mattson's applications of the principles elucidated by these experiments may seem to give strong support to the mutual precipitation theory. On closer examination, however, it is difficult to regard the mutual precipitation theory otherwise than as a helpful fiction. If mutual precipitation is to occur, then the positive sesquioxide sols and the negative silicic acid sols must first achieve separate existence. Can we imagine that, at one point in the weathering material, sesquioxide sols are formed, whilst at another point the corresponding silicic acid sols originate? The sesquioxides and silicic acid sols must at the moment of liberation be as closely adjacent as their positions in the parent lattice demand, and they can have no possibility of forming independent sols. Nevertheless, considering the isoelectric precipitate to be the most stable system in equilibrium with the liquid phase under the conditions of experiment, the assumption of a separate existence for the positive and negative sols may be dispensed with. The general agreement of Mattson's deductions with the observed facts under different conditions of weathering lends strong support to their general validity as a working hypothesis. Mattson's theories may be equally applicable to a theory of hydrolysis which conceives the first stage to consist in the formation of residual alumino-silicic and ferro-silicic acids.

GENERAL RESULTS OF HYDROLYSIS.

However we may conceive the process of hydrolysis, certain general consequences may be noted:—

(1) De-silicification occurs. Comparison of weathered materials with their parent rocks generally reveals a loss of silicic acid in the weathering process. The silicic acid is removed by percolating waters, to a great extent in the form of silicates of the alkaline and alkaline earths. It is significant that river waters, which contain the material lost by solution in the weathering processes, contain an excess of silicic acid over sesquioxides. In arid and semi-arid regions

incomplete leaching may prevent the removal of silicic acid liberated by hydrolysis. In such cases the weathering complex is characterised by a high content of silica.

(2) De-alkalisation occurs during hydrolysis. Comparison of fresh rocks with their weathered products reveals a general loss of calcium, magnesium, potassium, and sodium. Exceptions to this rule occur in arid regions, where alkali and alkaline earth salts may accumulate, giving rise to saline or alkaline soils.

The following data from G. E. Merrill⁴ illustrate the losses involved in chemical weathering (Table IV).

TABLE IV.—ANALYSES OF FRESH AND DECOMPOSED DIORITE. ALBEMARLE COUNTY, VIRGINIA (MERRILL).

(From Röcks, *Rock Weathering, and Soils*, by permission of The Macmillan Co., Publishers.)

	Fresh Diorite	Decomposed Diorite	CALCULATED	
			Per cent. Remaining	Per cent. Lost
SiO ₂	46.75	42.44	62.69	37.31
Al ₂ O ₃	17.61	25.51	100.00	0.00
Fe ₂ O ₃	16.79	19.20	78.97	21.03
CaO	9.46	0.37	2.70	97.30
MgO	5.12	0.21	2.83	97.17
K ₂ O	0.55	0.49	61.26	38.75
Na ₂ O	2.56	0.56	15.13	84.87
P ₂ O ₅	0.25	0.29	80.11	19.87
Ignition	0.92	10.92	100.00	0.00

Data such as those given above should be interpreted with caution since weathering processes vary considerably in intensity with the climate. Further, the differentiation of the weathered material in the horizons of a soil profile introduces an element of uncertainty as to what represents the immediate weathering product. In a podsol profile, the sesquioxides are leached out from the surface (A) horizons and deposited in lower (B) horizons. An estimate of the relative losses by weathering could only be formed from a consideration of weathered but undifferentiated material or of a composite sample of all horizons.

(3) Hydrolysis results in the formation of new substances, either by modification of the original minerals or by a partial re-synthesis of the products of decomposition. These substances are collectively referred to as the weathering complex, the zeolitic complex, the clay complex, or the inorganic soil colloid. They constitute the greater part of the clay fraction of mechanical analysis and are almost entirely found in that fraction. The study of the clay fraction, therefore, offers the most convenient basis for the elucidation of the composition and constitution of the weathering complex.

SOILS OF SECONDARY ORIGIN.

In our discussion of the weathering processes, we have given our attention principally to the origin of soils from the direct decomposition of crystalline rocks. Such soils may be termed primary soils. Many soils, and amongst them a considerable proportion of English soils, have not originated thus directly, but have been formed from deposits which have already passed through a cycle or cycles of weathering, transport, and consolidation. Such deposits as the Kimmeridge Clay and the Gault Clay have been formed from material which was produced by weathering in past ages, removed by the ordinary processes of denudation, and deposited as sub-aqueous sediments. It is obvious that the weathering complex of such soils is not to any great extent of contemporary origin. In the absence of heat metamorphism, leading to recrystallisation, we may regard all sedimentary rocks as containing the weathering complex formed in the epoch of their deposition. B. Polynov⁵ has suggested the descriptive, though cumbrous, terms *monochronogenous* and *heterochronogenous* for what we should prefer to term *primary* and *secondary* soils, respectively.

An examination of the clay fractions of soils of secondary origin reveals the fact that they are richer in silica than soils of primary origin. Thus the average molecular silica-sesquioxide ratio of the clay fraction of a number of N.

Welsh soils derived by primary weathering from crystalline rocks was found by G. W. Robinson⁶ to be 1.85, whilst the corresponding ratio for a number of soils derived from alluvia and other non-metamorphosed aqueous deposits was 2.67. This may be attributed to the concomitant precipitation of silicic acid from terrigenous waters at the time of their deposition. We have seen that hydrolysis involves a de-silicification of the parent minerals and that, with free percolation, this excess of silicic acid finds its way into the drainage and river system. It would thus appear that in the weathering cycle, the silica, separated from its original minerals in hydrolysis rejoins the weathering products during the stage of sub-aqueous deposition. From data obtained in the writer's laboratory it appears that the clay complex of calcareous sediments is considerably more siliceous than that of non-calcareous sediments. This is in accord with general experience that a high lime status is conservative of silicic acid in the clay complex.

DEGREE OF WEATHERING OF SOILS.

The degree of weathering is reflected in the character of the different soil fractions. At the one extreme are highly weathered soils of the humid tropics. In such soils, the only original minerals are those which are resistant to hydrolysis, such as quartz, magnetite, etc. Where the parent rock contains no unweatherable minerals, the resulting soil may consist entirely of the weathering complex. Some soils on analysis may yield up to 90% of clay. In other cases there is present a certain amount of concretionary material of secondary origin among the coarser fractions. Indeed, in some soils, the greater part of the soil may be represented by concretionary stones and gravel.

It has been remarked that an extreme laterite marks the ultimate end point of weathering and may be described as a dead soil, for the absence of silica implies the absence of a base absorbing complex and hence of bases. We should, however, hesitate before agreeing to regard laterite as the end

of the weathering process, seeing that it is probably formed by precipitation from sesquioxide sols and is not wholly a residual product in the strict sense of the term.

In the less weathered soils of temperate and cool regions there is generally a considerable proportion of weatherable minerals such as feldspars, micas, and ferromagnesian minerals in the silt and sand fractions. As a rule, soils of secondary origin are more highly weathered, i.e., contain less weatherable material, than soils of primary origin. This is due to the fact that they have already been submitted to one or more cycles of weathering. It is indeed quite possible that, even in Northern Europe, soils may occur which consist simply of quartz and the weathering complex.

An approximate indication of the degree of weathering of a soil may be obtained from the proportion of alumina in the clay fraction relative to the total amount of alumina present in the soil.* Since the principal weatherable silicates are all complex silicates of alumina and other bases, it follows that in a completely weathered soil, no alumina will be found in that portion of the soil from which the clay or weathering complex has been removed. Lateritic soils with concretionary alumina in the coarser fractions would, of course, be exceptions to this generalisation.

In soils where the weathering has been partly physical, the non-clay fraction contains unweathered aluminosilicate minerals, and the proportion of clay alumina to total alumina may give an approximate measure of the extent to which chemical weathering has proceeded.

Similar reasoning may be applied to the distribution of ferric oxide between clay and non-clay, but its applicability is limited by the possibility of the occurrence of unweatherable iron minerals such as magnetite and by the readiness with which hydrated ferric oxide forms coarse concretions.

The sesquioxides in the clay and in the total material of some typical soils were determined in the writer's laboratory, and the results are shown in Table V. The ratio of clay

*See G. W. Robinson and M. Richardson, *Nature*, 1932, **129**, pp. 581-582.

alumina to total alumina is given as a tentative measure of the degree of weathering. The corresponding ratio deduced from the ferric oxide figures is given with more reserve, for the reasons indicated¹ above.

TABLE V.—DEGREE OF WEATHERING OF SOILS AS SHOWN BY CONTENT OF ALUMINA AND FERRIC OXIDE IN CLAY AND TOTAL SOIL.

SOIL	Ignited Clay %	Per cent. Al_2O_3 in		Degree of Weathering %	Per cent. Fe_2O_3 in		Degree of Weathering %
		Clay	Total soil		Clay	Total soil	
Schistose subsoil, Wales	14.0	4.42	11.3	39.1	1.85	5.0	37.0
Shaly soil, Wales	21.0	7.05	17.5	40.3	2.55	7.6	33.5
Shaly sub-soil, Wales	19.5	6.5	19.7	33.0	2.3	7.4	31.1
Red loam, E. Africa	35.5	13.3	18.0	73.9	5.6	9.8	57.1
Yellow soil, India	20.2	5.35	9.6	55.7	2.1	3.7	57.0
Alluvial soil, Holland	48.5	12.85	13.7	93.4	2.96	6.1	48.5
Chalk soil, England	28.5	4.28	5.0	85.6	2.22	3.5	63.5
Red clay loam, E. Africa	50.0	17.05	19.0	89.7	8.25	8.7	94.3

The figures show the marked distinction between the relatively young Welsh soils, still in the early stages of chemical weathering, and the alluvial and chalk soils in which, presumably, the material has been submitted to one or more cycles of weathering. The Indian yellow soil and the African red loam occupy an intermediate position, whilst the E. African red clay loam is almost completely weathered. Like the Welsh soils, these are products of primary weathering, but the process has proceeded to a greater degree, particularly in the African red clay loam.

The data for ferric oxide generally confirm those for alumina, but the lower degree of weathering indicated suggests the presence of secondary ferric oxide or unweatherable iron minerals in the non-clay material.

Finally, it should be noted that even in the young Welsh soils, the fact that the degree of weathering is greater than the proportion of clay indicates a certain amount of chemical weathering. It would be interesting to find out whether there are soils in which no chemical weathering would be indicated by this method.

SOIL PROFILE DEVELOPMENT.

In considering the soil-forming processes, we have hitherto dealt only with those changes whereby rocks and their constituent minerals undergo physical and chemical weathering. We have now to examine a series of processes whereby, under the influence of water movements within the soil, the inorganic products of weathering and the organic matter undergo differentiation into definite layers or horizons, some characterised by impoverishment, and others by enrichment in certain constituents. The complete succession of horizons down to the undifferentiated parent material is called the *soil profile*.

Since the soil profile represents the result of all the soil forming processes, it is the natural unit of study. And therefore, in comparing soils for the purpose of classification, it is necessary to consider, not merely the surface soil, but all the horizons of the soil which constitute the profile. A classification of soils is a classification of soil profiles.

The differentiation into horizons has not always proceeded to its fullest expression possible under the local conditions of soil formation. Profiles which have not attained full development are termed *immature*. A soil profile may be immature where, as in the case of soils formed from recent alluvial or æolian deposits, time has not sufficed for mature development. In other cases erosion may have proceeded at such a rate as to keep pace with profile development and maintain a state of immaturity. This is the case on steep slopes under heavy rainfalls. Where, owing to interference with the natural cover of vegetation, erosion has intruded on a mature profile, a truncated profile characterised by the removal of certain horizons may result. Examples of truncated profiles are to be found in the South-Eastern United States, where deforestation by early settlers in a region subject to torrential rains and with a rolling topography has resulted in widespread erosion of the surface soil, with the result that over large areas the present cultivated soil represents a former sub-surface horizon.

Finally, even where erosion has not resulted in the prevention or destruction of soil profile development, cultivation may result in the confounding of the natural soil horizons. The typical agricultural soils of Britain show disturbed profiles. Apart from the interference with the natural course of profile development occasioned by the processes of cultivation and manuring, and the possibility of erosion after the removal of natural vegetation, the ordinary tillage operations cause a mixing of all the soil down to the limit of their action. And, therefore, if any vestiges of the original profile are to be found, they will occur below the depth affected by ordinary operations, i.e., about nine or ten inches.

Mature profiles are to be found, if at all, in areas of fairly level or gently rolling topography under natural vegetation, which have been for long ages subjected to the pedogenic processes characteristic of the locality. In some cases it may happen that profile development interrupted by erosion may be resumed on the establishment of a cover of natural vegetation. This leads to the formation of secondary profiles, such as may be observed in some of the uplands of Wales, where grassland or heath has been established in the place of former forest.

The development of a soil profile is the consequence of movements of water in the soil, and we may distinguish two extreme cases. Under humid conditions, there is an excess of rainfall over evaporation. There is thus a general tendency to downward movements of soil moisture, and the soil is subjected to a leaching process, whereby certain constituents are carried downwards and either deposited in lower horizons or completely removed in drainage. Under arid conditions, with an excess of potential evaporation over rainfall, such rain as falls moistens the soil to limited depths. After the cessation of rain, the soil moisture rises again to the surface under the influence of evaporation, with the result that translocation occurs alternately in both directions without any complete removal of constituents through leaching into the drainage. It should be added that evaporation is

not necessarily confined to the actual surface, for when the moisture content is reduced to a certain point, capillary rise is inappreciable. And thus, in the final stages of desiccation, the deposition of materials from solution may occur throughout an appreciable depth of the surface soil.

ELUVIATION.

We may refer to the translocation of material either mechanically or in solution, as *eluviation*, and two main types of eluviation may be distinguished, namely, (a) *mechanical eluviation*, in which, apart from any chemical differentiation, the finer fractions of the mineral portion of the soil are washed down to lower levels, and (b) *chemical eluviation* in which decomposition occurs and certain products thus liberated are translocated in true or colloidal solution, to be deposited in other horizons.

It is customary to refer to those horizons from which material has been removed by eluviation as the A or *eluviated* horizons, and to those horizons enriched by the deposition or precipitation of material from the eluviated horizons as the *illuvial* or B horizons. The undifferentiated parent material lying below is termed the C horizon.

Mechanical eluviation results in the development of a texture profile, characterised by a light-textured A horizon from which clay and the finer fractions have been removed, and an underlying heavy-textured B horizon enriched by the finer material eluviated from the A horizon. Strongly developed texture profiles are not common in England, but the increase in clay content in passing from soil to subsoil can generally be observed. In the South-Eastern United States well developed texture profiles are encountered. These show a sandy A horizon overlying a compact B horizon. It may be presumed that the presence of a considerable proportion of sand in the soil is a necessary condition for the development of a good texture profile, for a soil consisting mainly of fine material would not offer the necessary facility for water movements.

Whilst mechanical eluviation is governed principally by the intensity of the rainfall and the texture of the soil on which it falls, chemical eluviation involves more factors, and depends not only on the character of the parent material, but on the balance of rainfall and evaporation, the nature of the vegetation, and the temperature. The following are the principal groups of constituents affected by chemical eluviation, and the distinctive characters of soil profiles result from their differential movements in true or colloidal solution :—

- (a) Organic matter.
- (b) Silicic acid.
- (c) Hydrated alumina and hydrated ferric oxide.
- (d) Exchangeable bases associated with clay and humus.
- (e) Relatively insoluble salts, principally calcium sulphate and calcium carbonate.
- (f) Soluble salts, such as sodium sulphate, sodium chloride, etc.

The simplest type of eluviation is that affecting readily soluble salts such as sodium chloride and sodium sulphate. Under humid conditions with an excess of rainfall over evaporation, these, if formed, are completely leached out from the soil profile. Under conditions of incomplete leaching, their presence in the soil profile gives rise to a distinctive group of soils which will be described later under the heading of saline and alkaline soils.

Next in mobility to the salts of sodium and potassium are calcium sulphate and calcium carbonate. Certain important groups of soils are distinguished by the development of horizons in which calcium carbonate and calcium sulphate are deposited in definite horizons. They are, like the saline and alkaline soils, developed under conditions of incomplete leaching.

In soil profiles of more humid climates, calcium carbonate, calcium sulphate, and the salts of the alkali metals are completely leached out into the drainage. In addition,

there is also a leaching out of the exchangeable bases, which may proceed to such an extent that the colloidal complex loses its stability and undergoes destructive decomposition, accompanied by a greater or lesser degree of differentiation into its components. Thus, in soils of extremely humid climates, not only are the soluble salts and calcium carbonate completely leached out, but the bases combined with the complex acids of the colloidal complex become completely removed. Under these conditions the so-called humic acids become peptised as colloidal solutions and sink in the soil profile, where they become deposited as a humic B horizon. Under the same conditions, sesquioxides also become peptised, and are also leached from the A and deposited in the B horizon.

According to H. Harrassowitz⁷, sesquioxides, under alternating conditions of humidity, may be carried to the surface as humus-protected sols and there undergo irreversible precipitation. A. Reifenberg⁸ postulates a similar process with colloidal silicic acid as the protective colloid. Under such conditions, superficial deposits of sesquioxides may be regarded as constituting illuvial horizons.

Eluviation does not consist necessarily in the downward translocation of material. Indeed, even where the net movement is in a downward direction, as in the pod sols, the actual deposition in the B horizon may take place from ascending solutions during periods of drought—a type of illuvial deposition which is even more pronounced in lateritic profiles. In those portions of the humid tropics which are distinguished by well marked wet and dry periods, mature soil profiles always contain concretionary material irreversibly deposited during dry periods from relatively concentrated soil solutions. This concretionary material varies in development from small scattered pisoliths to the massive crusts of the extreme laterites.

Surface incrustations of calcium carbonate are encountered in semi-desert regions as, for example, in the South African Karroo.

The presence near the surface of ground water may lead to the development of a characteristic type of profile—the “*gley*” profiles of the mid-European writers. The alternation of oxidising and reducing conditions in the zone of fluctuation of the water table produces the rusty mottlings of the “*gley*” horizon. The changes occurring under submerged conditions are considered at a later point.

The character of the natural vegetation is reflected in the organic matter profile. Under coniferous forest or heath, the residues of vegetation accumulate as an acid peaty layer sharply distinguished from the mineral subsoil. Permanently wet conditions at the surface result in peat development. Under deciduous forest, plant residues, including leaf-fall and ground vegetation, are incorporated with the soil as “mild” humus, the dissemination through the soil being facilitated by the activity of earth worms. Under steppe, prairie, or savanna conditions, there is also a uniform distribution of organic matter corresponding with the root range of natural vegetation.

Finally, in addition to the texture profile and the mineral and organic profiles, there is the development of a structure profile. We have seen that soil occurs naturally in the form of compound aggregates or crumbs formed by the ultimate particles. The character of the pedogenic processes is often reflected in the type of compound aggregate formed. Thus, in a podsol profile, the superficial horizon tends to have single grain structure, becoming dusty on drying. The illuvial or B horizon tends to be compacted into a pan. In the tshernosems or black earths, the surface horizon has a porous granular structure, whilst the lower horizons may exhibit nut structure or columnar structure.

It cannot be too strongly emphasised that the soil profile, since it gives the summation of all pedogenic factors, including parent material, climate, vegetation and soil history, is the natural unit of soil study. Even where the natural profile has been partially obliterated by cultivation or truncated by erosion, it is often possible to infer, from the

profile thus modified, -the character of the original profile. The general tendency of profile development can also be observed in soils which have not yet attained their fullest development.

We have, for convenience, considered the processes of weathering and eluviation separately. In some cases, doubtless, such a chronological succession does occur, particularly in soils of secondary origin. In the case of soils of primary weathering, it is scarcely possible to postulate such a separation of the processes. Where a podsol profile is developed from a crystalline rock we may assume that the processes of physical and chemical weathering and the processes of eluviation proceed concomitantly. In the case of laterites, also, the characteristic features of the profile are recognisable in the earliest stages of development.

There are, however, large areas, including those localities of Central Europe and Russia where modern profile study has originated, in which the differentiation of horizons implied by profile development has taken place in material which has already undergone weathering and which has been transported from its original site by ice, water, or wind. It may be assumed, even in these cases, that weathering also proceeds concomitantly with profile development. //

PAN FORMATION.

Under certain conditions materials existing in true or colloidal solution in the soil moisture may be irreversibly precipitated in definite horizons whose constituent particles thereby become cemented together to form pans which in some cases are impervious alike to root growth and to the movements of water.

The most widely spread type of pan is that formed by deposition of hydrated ferric oxide, the so-called iron pan or "*ortstein*." The presence of this material as a cementing agent is generally evident from the colour. In many cases, however, the ferric oxide is accompanied by alumina. So far as the writer is aware there are no pans formed by

alumina alone as a cementing agent. Iron pans are formed in practically all latitudes. They occur in association with the podsoles under northern coniferous forests and heaths, and in many tropical and desert soils.

It is possible to distinguish a number of types of sesquioxide pan formation. In the first place, there is the development of iron pan under podsol conditions in the presence of free drainage through the soil profile. Whilst the mobilising of sesquioxides under the protective action of humus sols is readily explainable, it is not so easy to arrive at a satisfactory explanation of the redeposition of sesquioxides in an illuvial horizon. One theory is that precipitation is due to the reaction between the sesquioxide sols and the inorganic material of the horizon of precipitation, which has a higher base status than the horizon from which the sesquioxides have been leached. It does not appear, however, that there is any marked difference in base status between the eluviated and the illuvial horizons in podsol profiles. Such differences as occur, however, do show a slight increase in basicity in passing from the A to the B horizon.

Another possibility is that the original tendency to pan formation is given by a restriction of freedom of drainage by the mechanical deposition of fine material in the B horizon. Whilst electronegative sols can pass through fine capillaries, electropositive sols such as those of ferric oxide and alumina are precipitated. It appears, however, that the sols in the soil are always electronegative.

A more probable theory of the formation of sesquioxide pans is that they represent irreversible precipitations from ascending currents of soil moisture during periods of drought. During the wetter months of the year—in our climate during the winter months—the general movement of water in the soil is downward. During dry periods the surface soil becomes dry and there is an upward movement of moisture from below. Evaporation takes place at the surface and even although complete desiccation may not extend more than a few centimetres into the soil, the fall in moisture

content in the soil may cause a sufficient increase in concentration of the dissolved or peptised constituents for precipitation to occur.

Another type of sesquioxide pan formation is encountered in water-logged soils in the zone through which the level of the water table fluctuates. Under the reducing conditions which obtain below the water table, ferric compounds undergo reduction and such compounds as ferrous sulphide and ferrous carbonate are produced. They are readily oxidised during periods of low water table. Further ferrous compounds in true or colloidal solution are brought into contact with the air during periods of capillary rise from the water table. The deposits formed in this way are principally hydrated ferric oxides since compounds of iron are more mobile than aluminium compounds. In some cases the deposition may reach such proportions as to yield bog iron ore. In certain types of tropical soils, with impeded drainage, the deposition of iron oxides may result in the development of highly indurated concretionary material, known in Africa as "murram" and in Cuba as "mocarrero."

The presence of considerable proportions of stones or pebbles appears to favour the formation of hard pans.

The most extreme instance of pan formation is the development of ferruginous and aluminous crusts in the surface horizons of laterite profiles. These will be described in detail in a later chapter. It should be noted that laterite crusts are associated with tropical climates in which alternations of rainfall and drought occur. They are not formed in tropical regions of unbroken humidity. The development of deposits of sesquioxides varies considerably. Where the tendency to concretionary formation is less strongly developed, pisoliths may occur in the upper horizons. These consist mainly of hydrated sesquioxides, but silica may also be present.

In a succession of soils starting with tropical rain forest and passing to less humid regions, concretionary material

begins to be observed under monsoon forest and becomes more developed in savanna, reaching its greatest development in certain types of laterite, characteristic of this formation. In the direction of increasing aridity, sesquioxide concretions become less evident and calcareous pans become increasingly developed as steppe and desert are reached.

Calcareous and gypseous pans are essentially formations of semi-arid and arid climates. In the black earths and their related types, calcium carbonate and calcium sulphate are deposited in well marked horizons, which approach the surface more nearly the more arid is the climate. The deposits of calcium carbonate and gypsum in black earth profiles are generally of an intermittent character and are scarcely to be considered as pans in the strict sense. In desert and semi-desert regions such as the South African Karroo, massive rock-like deposits of calcium carbonate may occur at or near the surface. Calcareous pans may also occur under conditions of restricted drainage, as in certain tropical soils.

CHANGES OCCURRING UNDER IMPEDED DRAINAGE OR SUBMERGED CONDITIONS.

It frequently happens that either through the presence of a high water table or through the occurrence of impervious strata, certain soil horizons are submitted permanently or intermittently to complete saturation. Such conditions result in changes which, from the frequency of their operation, must be reckoned among the most important of the pedogenic processes.

The most notable consequence of excessive moisture is the exclusion of air and the restriction of free percolation. The first result of the exclusion of air is the limitation of root development to the immediate surface and the encouragement of a type of natural vegetation adapted to these conditions. In this way, marsh, swamp, or bog associations are developed. Where the soil is continuously waterlogged to the surface, peat develops by anaerobic humifi-

cation; in the early stages under the influences of ground-water, but later, where the rainfall is sufficient, under the influence of atmospheric moisture alone. Even in cases where the ground-water level is not sufficiently near the surface for peat development, there is generally a development of peaty organic matter, and a sharp change in organic matter content is noted in passing from the humic layer to the horizon in which root development is inhibited by lack of air.

The second consequence of exclusion of air through waterlogging is the prevalence of processes of reduction. Ferric compounds are reduced to ferrous compounds, as shown by the grey or greenish-grey mineral colour of waterlogged horizons and the occurrence of such compounds as ferrous sulphide, iron pyrites, marcasite, and vivianite. Where there is a fluctuation of the ground-water level the alternation of oxidising and reducing conditions leads to a characteristic brown or brownish-yellow mottling. In extreme cases notable deposits of hydrated ferric oxide may be formed, represented by the bog iron ore of northern countries and the "murrum" of the tropics. Manganese dioxide is frequently deposited near ground-water horizons.

The restriction of percolation hinders the removal of products of silicate weathering. An important consequence of this is that the weathering complex formed under such conditions is markedly more siliceous than that formed under conditions where complete leaching is possible. This, in an equal measure with reducing conditions, contributes to the production of soils with grey mineral colours, as distinct from the reddish or brownish soils which result when the removal of excess silicic acid is unchecked.

In basin shaped areas where waterlogging is accompanied by high evaporation, soluble salts, including calcium carbonate, gypsum, and sodium salts, may accumulate, giving rise in some cases to gypseous or calcareous deposits or concretions, and in other cases producing saline or alkaline soils.

The incomplete removal of soluble materials, except in the more humid regions, results in a relatively high base status. Under such conditions the humification of organic residues appears to produce a dark type of humus. Such soils as the black turf soils of S. Africa may have dark grey or black colours and yet contain less than 1% of organic carbon. The black colour of certain prairie soils in N. America developed in association with a high water table or impeded drainage may also be noted.

W. O. Robinson⁹ found that, in the presence of organic matter, submergence is followed by a rapid increase in the solubility of iron, manganese, calcium and magnesium. Solution is effected by the presence of carbon dioxide from the microbiological decomposition of organic matter, which can take place with considerable rapidity. The iron comes into solution in the ferrous state, probably as ferrous hydrogen carbonate. Where leaching is possible, considerable losses of iron, manganese, calcium, and magnesium may occur. Where organic matter is absent, or present in small amount, the solubilisation of iron, manganese, calcium, and magnesium is negligible. Any acidity directly resulting from submergence is attributable to the carbon dioxide formed by decomposition of organic matter. The formation of soluble carbonates must, however, exert an alkalinising effect. The removal by leaching of the bases thus mobilised would cause a lowering of the base status.

SOIL EROSION.

By soil erosion is understood the removal of material from the surface of the soil by the agency of running water, wind, or even by gravity. Since the superficial layers of the soil are richest in plant food and are the feeding ground of plant roots, this process involves a definite loss of valuable material and, if it becomes sufficiently intense, may lead to the complete destruction of the soil as the seat of plant life.

Under the most stable conditions, there is always a certain loss of material from the soil surface, but this is more

or less balanced by the formation of fresh soil through the processes of weathering and the accumulation of organic matter from the residues of vegetation. Such a balance between erosion and soil formation permits the development of a mature soil profile and is realised under a stable closed vegetation with flat or gently rolling topography. Under these conditions wastage is comparatively slight and may amount, when reckoned on the whole land surface, to less than a foot per century. A considerable proportion of this wastage represents material lost by solution in percolating waters which find their way into rivers and ultimately to the sea.

It is obvious that, where considerable erosion occurs, the natural soil profile is destroyed or truncated and, indeed, may never attain full development. In considering the relationships of the soils of a particular region to the great soil groups, it is therefore necessary to make allowance for the possible effects of erosion in the past. This is particularly the case in humid tropical regions, where, as will be seen, erosion is intense and has led to considerable modifications in profile development.

Considering in the first place erosion by running water, we may distinguish two types, namely, *sheet erosion* operating over whole areas, and *gullying*, in which the erosion is confined to comparatively narrow channels. In the first type of erosion, although the results are rarely spectacular, its operation over large areas leads to enormous losses of material, which either accumulates in valleys or hollows or else is carried away by rivers to the coast. In the second type, the effects of erosion are evident in the deep channels which are cut; but the damage due to the loss of material is less serious than the resultant lowering of the water table, which leads to excessive drainage, and damage to vegetation through drought. Examples of this kind of erosion are common in S. Africa and in parts of the United States.

The rate of erosion of soil by running water depends principally on four factors, namely, (1) the amount and

intensity of the rainfall; (2) the angle of slope; (3) the presence or absence of a vegetative cover, and, if present, its character; and (4) the nature of the soil itself.

The actual amount of annual rainfall is of less importance than its distribution and the extent to which it occurs in torrential downpours. It is only when the soil is unable to dispose of rainfall by percolation that running water capable of producing erosion appears on the surface. Even in the wettest districts of Britain, it is only in exceptional cases that rain falls so quickly that the soil is unable to dispose of it by percolation. The position is completely different under humid tropical and sub-tropical conditions. Here the rain occurs to a large extent in intense falls of comparatively short duration—as much as six inches per hour having been recorded—with the result that running water soon appears on the surface. Similar conditions occur in certain arid regions where the scanty precipitation may occur mainly in the form of occasional torrential downpours. It is a remarkable circumstance that, in arid regions such as Nevada and Southern California, evidences of erosion and transport of material by water action are more obvious than in more humid climates such as that of Britain.

It is obvious that the slope of the land has an important bearing on its liability to erosion. Other things being equal, the amount of *run-off* and therefore erosion is more intense on steep slopes. R. E. Dickson¹⁰ obtained the following figures for *run-off* in inches from a Texas clay soil:—

Gradient	Level	1 in 100	1 in 50
Run-off water ...	0.62	3.42	3.80

H. H. Bennett¹¹ reports a 25% increase in erosion for each 1% increase in gradient over 1 in 27 on a loam soil in Missouri.

Even with the comparatively slight erosion to which British soils are subjected, the accumulation of soil above fences running along contours can be generally observed, and is the more evident the steeper the slope.

The importance of a cover of vegetation in checking erosion is obvious. Not only does it protect the soil from the beating action of rain, but the network of roots and stems hinders erosion even when water appears as run-off. Further, the presence of a closed vegetative cover assists in maintaining the top soil in such a physical condition that it can absorb intense downpours without suffering the mechanical deflocculation which prevents free percolation.

In regions which, from the amount and nature of the rainfall, are liable to erosion, the destruction of the natural vegetation is fraught with serious consequences unless carefully planned measures are taken to minimise the danger. Unfortunately, over large areas of the tropics and sub-tropics, irreparable damage has already taken place, resulting in the complete removal of the surface soil and the exposure of raw subsoils or even bare rock.

The destruction of the natural vegetative cover by grazing animals may expose soil to the influences of erosion. This has occurred in many parts of the South African Veldt,¹² where the "kraaling" of stock, rendered necessary on account of the menace of jackals, has resulted in intensive trampling, followed by "slooting" (gullyng). The consequent lowering of the water table has intensified the effects of drought and led to a general lowering of stock-carrying capacity.

Soils laid bare by arable cultivation or by the agency of grazing animals are naturally subject to erosion; but important differences may be remarked in their susceptibility to this process. Generally speaking, it may be said that a granular or crumb structure favours resistance to erosion, whilst a deflocculated or single grain structure renders soil liable to erosion. The presence of a high proportion of organic matter is helpful in diminishing liability to erosion. It is noteworthy that, even on the steepest Welsh hillsides in cultivation, erosion is almost imperceptible, a circumstance which may be attributed to the high proportions of organic matter present in the soil.

H. H. Bennett¹¹ has shown that in the South-Eastern United States, Cuba, and Central America, soils fall into two classes with respect to their liability to erosion. On the one hand, soils whose clay constituent is characterised by excess of sesquioxides are markedly resistant to erosion, probably owing to their tendency to assume a crumb structure which favours the rapid absorption of rainfall. On the other hand are soils whose clay is of a more siliceous type. These are more readily eroded owing to their tendency to deflocculation.

The extent of erosion in the tropics and certain parts of the sub-tropics is scarcely apprehended by those who are only familiar with British conditions. A measure of its intensity is furnished by the content of suspended matter in river waters. Whilst in the rivers of Western Europe this amount is generally less than fifty grams per cubic metre, from 2,000 to 10,000 grams per cubic metre are found in tropical rivers. From South Africa, silt contents of over 2% are reported.

Bennett (*loc. cit.*) has estimated that, in the Piedmont region of the United States of America, about 65% of the area which has been or is now under cultivation has lost from 4 to 18 inches of soil, and that more than 50% of the alluvial land formerly cultivated in valley bottoms has been covered up by erosional débris. This process dates from the settlement and deforestation of the area by the white settlers. Profiles are frequently truncated by the removal of the A and even part of the B horizon; so that a large proportion of the area at present under cultivation represents what was originally B horizon material. An appreciation of the extent of the erosion is necessary in order to understand the soil profile relationships of this region, where the assumption that the surface soil represents an A horizon may frequently lead the investigator into serious error.

It may be conjectured that erosion, analogous to that which has proceeded in the Eastern United States within the last three centuries, has taken place also in Great

Britain, although at an earlier period, and, in view of the smaller prevalence of torrential rain, to a smaller degree. Evidence for this in Wales is furnished by the observations of G. W. Robinson¹³ on the changes in composition of the clay fraction in the soil profile. Whilst the general tendency under Welsh climatic conditions is towards a leaching out of sesquioxides from the surface horizons, many of the surface soils show the high proportions of sesquioxides associated with B horizons, and this tendency is most marked in situations where rainfall and topography might be expected to have resulted in heavy erosion. It may be conjectured that considerable erosion took place when the land was first deforested, with consequent removal of material of a relatively siliceous character, such as is now commonly found in hollows and bottoms. With the institution of grass husbandry and the consequent increase in organic matter, erosion became checked and, if we may draw a parallel from Bennett's observations in America, the sesquioxidic character of the clay probably contributed to stability in this respect. In the higher horizons, the grass vegetation gradually degenerated to heath, and secondary podsoles were developed in the primitive B horizon.

The evidence from England is less plain, but in many districts, notably in chalk country, stretches of colluvial soils on slopes, and shallow profiles in the uplands suggest past erosion.

The effects of erosion by wind are more plainly seen in the accumulation of transported material than in the actual denudation which accompanies the process. The great wind-borne loess deposits are the result of wind erosion acting in past ages.

Wind erosion, at the present day, is principally active in desert and semi-desert regions. Since only the finest particles are liable to transport by wind it frequently happens that a sorting out occurs whereby fine material is removed and the coarser gravel and stones remain behind as "desert pavement." This effect is only operative at the

immediate surface and finer grained material is found below the level of stones and gravel.

The occasional occurrence of hard layers of calcium carbonate at the surface in parts of the South African Karroo and other deserts may be due to removal of formerly overlying layers by wind action.

NATURAL VEGETATION AS A PEDOGENIC FACTOR.

The connexion between vegetation and soil is so intimate that the principal soil groups can be almost completely defined according to the types of vegetation under which they occur. Broadly speaking, the character of the natural vegetation expresses the summation of the climatic factors under which it grows. Yet there are plant associations whose occurrence is determined mainly by soil conditions, whilst in other cases, climatic factors, though dominant, may be modified by the character of the soil. For example, forest is generally found under humid conditions. On the one hand, extreme humidity leads to the replacement of forest by bog or moor, whilst, on the other hand, with decrease of humidity, transition to steppe or savanna occurs. But, even in a region of typical forest climate, local factors of topography and drainage may lead to the development of bog or moor vegetation. Similarly, in a steppe region, local conditions may lead to the development of forest.

The effect of soil conditions is seen most markedly in the transition zone between vegetation types, and may cause markedly different plant associations to occur under identical climates. An example of this is the former occurrence of steppe vegetation, now represented by black earths, in the forest region of Western Europe.

And thus, whilst it is generally permissible to regard natural vegetation among the factors in soil formation, the reciprocal character of the contract should not be overlooked. Nor can vegetation be regarded as an independent factor, seeing that it is itself controlled by climate. The operation

of climate as a pedogenic factor is both direct on the soil itself and indirect through the vegetation.

The effect of forest as a vegetative cover is exerted mainly through the organic matter of the soil on which it occurs, and depends on the nature of the leaf-fall and the presence or absence of ground vegetation. In northern coniferous forests there is practically no ground vegetation, and the only source of soil organic matter is the leaf-fall, which, being naturally poor in mineral substances, tends to form a superficial layer of acid peaty humus, favourable to the podsolising processes. There are, however, differences among the conifers with respect to the mineral status of their leaf-fall.

Under deciduous forest conditions are markedly different. Not only is the leaf-fall richer in mineral constituents, but there is generally a considerable contribution of organic matter from ground vegetation. The relatively high base status of the litter and the residues from ground vegetation contribute to the formation of a neutral type of humus, whilst the presence of earth worms ensures its distribution throughout the profile. NB

The organic matter profile of coniferous forest soils thus differs from that of deciduous forest soils. In the one case there is a sharply defined humic layer with a strongly acid reaction overlying a mineral soil which is relatively poor in organic matter. In the other case there is a moderate content of neutral or only slightly acid organic matter which decreases gradually in amount from the surface.

Premising that the excessive development of peaty humus is disadvantageous to tree growth, even in coniferous forests, the importance of a clear view of the part played by soil, climate, and vegetation, is apparent. A soil developed from parent materials such as quartzose sand, or acid igneous rocks, can afford little compensation for the low mineral status of the leaf-fall of trees such as *Pinus sylvestris* and *Picea excelsa*, particularly if, owing to high rainfall and low mean temperatures, leaching is intense. Under such

conditions, the best prospect of checking undue development of acid humus would be through the growth of trees whose leaf-fall has naturally a higher mineral status and which would therefore tend to conserve in the surface layer such base supplies as the soil possesses. It is possible that the encouragement of a ground vegetation by relatively open planting might also help to counter the tendency to acid humus development.

Under grassland or steppe conditions, the organic matter status of the soil is appreciably higher than under forest. This leads to the apparently paradoxical result that in passing from a forest climate to the drier climate of steppe or prairie there is a marked increase in organic matter, which is maintained until the maximum content is reached in the middle of the black earth belt. Thereafter, increasing aridity, by its effect in curtailing plant growth, is reflected in a decrease in organic matter.

The higher organic matter content of steppe soils compared with forest soils is due to a number of circumstances. Firstly, under forest, accessions of organic matter are mainly in the form of leaf-fall which, accumulating on the surface, is exposed to serious losses by oxidative decomposition. Under steppe conditions, plant residues are added both at the surface, from the remains of leaves and stems of plants, and also within the soil, from dead plant roots. Secondly, the activity of earth worms and other burrowing animals is greater under steppe than under forest. This ensures a more rapid and thorough incorporation of residues throughout the profile. Thirdly, summer drought under steppe conditions results in a restriction of biological decomposition to an extent not realised under the more humid forest conditions.

BIBLIOGRAPHY.

- ¹ VAN HISE, C. R., "A treatise on metamorphism." *U.S. Geol. Survey Bull.*, 1904.
- ² CLARKE, F. W., "Data of geo-chemistry." *U.S. Geol. Survey Bull.*, p. 770.

- ³ VAN BEMMELEN, J. M., "Die Absorptionsverbindungen und die Absorptionsvermögen der Ackererde." *Landw. Vers. Stat.*, 1888, **35**, pp. 67-136.
- ⁴ MERRILL, G. E., *Rocks, rock-weathering, and soils*. New York, 1913, p. 207.
- ⁵ POLYNOV, B., "Das Muttergesteine als Faktor der Bodenbildung und als Kriterium für die Bodenklassifikation." *Soil Research*, 1930, **2**, pp. 165-180.
- ⁶ ROBINSON, G. W., "The nature of clay and its significance in the weathering cycle." *Nature*, 1928, **121**, p. 903.
- ⁷ HARRASSOWITZ, H., "Laterit, Material und Versuch erdgeschichtlicher Auswertung." *Fortschr. Geol. Palæontol.*, 1926, **4**, pp. 253-566.
- ⁸ REIFENBERG, A., "Entstehung der Mediterran-Roterde." *Kolloid. Beih.*, 1929, **28**, pp. 1-93.
- ⁹ ROBINSON, W. O., "Changes occurring in submerged soils." *Soil Sci.*, 1930, **30**, pp. 197-217.
- ¹⁰ DICKSON, R. E., "The results and significance of the Spur (Texas) run-off and erosion experiments." *J. Amer. Soc. Agronomy*, 1929, **21**, pp. 415-422.
- ¹¹ BENNETT, H. H., "Some aspects of soil erosion as a national problem." *Rep. 9th Meeting Amer. Soil Survey Assoc.*, 1929, pp. 55-74.
- ¹² Union of S. Africa, *Final Report of Drought Investigation Commission, S. Africa*, 1923.
- ¹³ ROBINSON, G. W., "The development of the soil profile in N. Wales as illustrated by the character of the clay fraction." *J. Agric. Sci.*, 1930, **20**, 618-639.

CHAPTER IV.

THE CLAY COMPLEX.

RELATIONSHIP TO WEATHERING COMPLEX AND CLAY FRACTION.

WE have seen that the soil may be conveniently considered as being composed of a relatively inert framework of unweathered minerals, and a colloidal complex consisting mainly of the products of the chemical weathering of silicates, together with humus, which is the product of the decay of plant and animal residues in the soil. By reason of the intimate association of the inorganic and the organic colloidal material in the soil, it is convenient for many purposes to consider them as a single constituent—the colloidal complex. But in view of the importance of the inorganic colloidal material in the weathering cycle and in the processes of eluviation, which, by their differing operation, distinguish the great soil groups, we shall in the present chapter confine our attention mainly to the clay complex.

In considering the mechanical composition of the soil, we had occasion to distinguish the so-called clay fraction, which was defined in terms of particle size or, more correctly, settling-velocity. We may also recognise an inorganic colloidal fraction, or complex, which is the reactive mineral constituent of the soil and which confers on it those physical properties which differentiate soil from physically comminuted powders. Finally, we recognise a certain portion of the soil as consisting of material of secondary origin which has resulted from the chemical weathering of silicate minerals. Can we regard the three groups thus defined as being identical? In other words, can we assume that the

clay fraction contains all the chemically reactive colloidal inorganic material of the soil, and that this is identical with the fraction which has resulted from the chemical weathering of mineral silicates?

Whilst we cannot assume the complete identity of these groups, there is probably no great error, for ordinary soils, in the assumption. Certain tropical soils may, indeed, contain material of secondary origin in the fractions other than the clay. In lateritic soils, the coarse material may be almost entirely secondary. Coarse concretionary material may also occur in soils developed in proximity to a water table. But such secondary material, though strictly belonging to the weathering complex, would scarcely exhibit chemical reactivity or confer on the soil the characteristic physical properties of colloids. Apart from such cases as these, the material coarser than clay is generally composed of original crystalline unweathered minerals, and rock fragments. We may assume, therefore, with the exceptions mentioned, that the clay fraction includes all the secondary products of weathering—the weathering complex.

It is less easy to satisfy ourselves that the clay fraction as defined in mechanical analysis (<0.002 mm.) does not contain unweathered minerals. Crystalline material is certainly present. The curious satin-like appearance sometimes observed in agitated suspensions of clay fraction is due to reflection from small platelets. X-ray analysis has also revealed the ultimate crystalline character of the colloidal constituent of many clays. This micro-crystalline material is, however, of secondary origin, and we are more concerned to know whether original unweathered minerals are present.

A. F. Joseph¹ found that the whole of the clay fraction isolated from certain Sudan soils could be further dispersed into "ultra clay" and was presumably of uniform character. S. B. Hendricks and W. H. Fry² found no felspar or mica below 0.001 mm. (1μ). R. Bradfield³ found no essential difference between suspensoid clay (0.4 – 1.2μ) and emulsoid clay ($<0.1\mu$) from a Missouri soil.

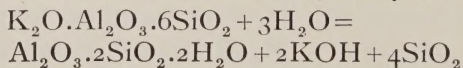
On the other hand, A. Atterberg⁴ concluded that biotite forms the plastic ingredient of certain northern clays, whilst hæmatite and limonite are present in the clay of terra rossa. He found that other lamellar minerals, such as kaolinite, talc, and muscovite, can exhibit the clay property of plasticity when sufficiently comminuted.

From a consideration of mechanical composition curves of physically comminuted materials, G. W. Robinson⁵ concluded the lower limit of physical subdivision to be represented by particles of approximately 0.0006 mm. diameter ($\log V = 5.0$). Data for the composition of successive fractions of typical soils also suggest that, except in certain highly weathered tropical soils, unweathered minerals are present in the fractions above 0.0006 mm. in diameter. It seems possible, therefore, that the clay fraction as ordinarily defined may contain some unweathered physically comminuted minerals in the upper part of its range, but the proportions thus present would generally be small. Although, in the writer's view, it would have been better to fix the clay limit lower than 0.002 mm., there is probably no serious error, for most purposes, in identifying the clay fraction with the weathering complex. The error is likely to be most serious in certain soils in which the weathering has been mainly physical.

EARLIER VIEWS ON THE NATURE OF CLAY.

Premising that the clay complex is generally and mainly the product of the chemical weathering of silicate minerals, its constitution has been variously pictured.

The earliest theory, still extant in many textbooks, regards kaolinite as the essential mineral of clay, and chemical weathering as the process of kaolinisation, represented ideally, in the case of orthoclase felspar, by the equation :—



It should be remarked, however, that kaolin, whose characteristic mineral is kaolinite, is the product of deep-

seated hydrothermal decomposition, which differs entirely from the epigene processes whereby soils are formed. Further, it is lacking in the properties of plasticity, colloidal imbibition, and base exchange, which are characteristic of soils. The kaolinisation theory is thus inadequate as a general explanation of the origin of the clay complex, although the clay complex may contain kaolinite or minerals similar in constitution to kaolinite, formed by epigene processes.

J. M. Van Bemmelen⁶ considered the weathering complex of the soil to consist of two fractions, namely, a complex A, soluble in hot concentrated hydrochloric acid, and having a molecular $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratio varying from 3 to 6, and a complex B, soluble only in hot concentrated sulphuric acid, having a molecular $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratio varying from 2 to 3. The complex A was considered to be an absorption compound of indefinite composition and colloidal character, whilst complex B was considered to be essentially similar to kaolin and to be less reactive than complex A.

H. Stremme⁷ distinguishes three groups of hydrous aluminium silicates formed by the weathering of mineral silicates. The first group are residual products from the decomposition of feldspars and feldspathoids, and approximate in composition to kaolinite, $\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 2\text{H}_2\text{O}$. Though not demonstrably crystalline, they do not form gels. The members of the second group, the allophanoids, have $\text{SiO}_2/\text{Al}_2\text{O}_3$ ratios varying from less than 2.0 to greater than 2.0. By replacement of aluminium with iron, corresponding ferric silicates are formed such as nontronite, chloropal, etc. They are typical gels and are formed partly as precipitates and partly as residual products from the decomposition of ferromagnesian minerals. H. Harrassowitz⁸ considers them to originate also from lime feldspars. The third group includes the zeolites, which do not apparently occur in soils, but represent the crystalline prototypes of the allophanoids.

R. Ganssen⁹, from the results of acid extraction designed to separate the weathering complex from the un-

weathered portion of the soil, postulates a compound of the formula, $\text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2 \cdot \text{RO}(\text{R}_2\text{O})$, and accounts for actual variations from this formula by the presence of excess of silica or excess of alumina. It may be doubted, however, if any method based on acid extraction can effect a separation of the clay complex from the unweathered portion of the soil, for certain minerals have an appreciable solubility even in dilute acids. For example, a powdered quartz-diorite, in which, of course, no weathered constituent was present, was found to lose 10.8% silica, 20.6% alumina and ferric oxide through extraction with 20% hydrochloric acid at 100°C.

A defect of the earlier theories as to the constitution of the clay or weathering complex of soils is that they were generally conceived in terms of compounds or complexes built up essentially of silica and alumina, with iron oxides as adventitious constituents. It is now known that iron compounds form essential ingredients in the clay complex, for clays containing notable proportions of iron occur in which the presence of adventitious ferric oxide cannot be assumed. Any theory of the nature of the clay complex must take account of the possible existence of complex ferro- or ferri-silicates and of the possibility of isomorphous replacement of aluminium by iron in complex alumino-silicates.

The similarity of the base exchange reactions of the soil to those exhibited by zeolites has led certain workers to postulate the presence of these minerals in the weathering complex. Although none of the known zeolites have been shown to be present in soils, the clay complex has frequently been termed the zeolithic (or zeolitic) complex.

COMPOSITION OF THE CLAY FRACTION.

During recent years considerable light has been shown on the constitution of the weathering complex by the examination of the clay fraction separated in mechanical analysis. Some data for the chemical composition of the clay fraction of typical soils are given in Table VI. It should be observed that, whilst in most cases the fraction <0.002 mm.

has been used, the data of W. O. Robinson and R. S. Holmes relate to the ultra-clay ($<0.3\mu$).

TABLE VI.—COMPOSITION OF CLAY FRACTIONS OF SOILS.

SOIL	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	TiO ₂	CaO	MgO	K ₂ O	Na ₂ O	Authority
I.	44.89	22.59	7.75	0.47	1.48	1.44	1.36	0.22	} W. O. Robinson and R. S. Holmes ¹⁰
II.	33.95	36.06	11.02	0.62	0.31	0.40	0.56	0.44	
III.	53.34	20.77	7.15	0.81	5.01	1.39	n d	n d.	
IV.	49.5	30.9	12.2	n d.	n d.	n d.	n d.	n d.	A. D. Hall ¹¹
V.	44.08	27.64	21.81	n d.	0.58	1.61	1.10	0.96	J. Hendrick and W. G. Ogg ¹²
VI.	43.9	35.6	19.9	n d.	0.3	0.4	0.8	n d.	} B. de C. Marchand and C. R. Van der Merwe ¹³
VII.	58.9	17.3	20.8	n d.	0.2	2.4	1.2	n d.	
VIII.	46.6	33.3	12.6	n d.	n d.	n d.	n d.	n d.	G. W. Robinson ¹⁴
IX.	10.19	15.84	62.51	0.20	0.23	0.05	tr.	tr.	G. Edginton ¹⁵
X.	27.6	37.8	26.8	1.4	n d.	n d.	n d.	n d.	G. W. Robinson ¹⁴

- I. Carrington Loam, 0-9", Iowa, U.S.A.
- II. Cecil Clay, 0-9", Georgia, U.S.A.
- III. Houston Black Clay, 0-12", Texas, U.S.A.
- IV. Average of 8 soils from Rothamsted, England. Ignited.
- V. Craibstone, Scotland. Ignited.
- VI. Transvaal Red Loam. Ignited.
- VII. Transvaal Black Turf. Ignited.
- VIII. Shale Soil, N. Wales. Ignited.
- IX. Nipe Lateritic Clay, Cuba.
- X. Red Loam, Tanganyika. Ignited.

It will be seen that the composition of the clay shows a very wide variation, and that in all cases it departs from the composition of kaolin, once supposed to be the characteristic ingredient of clay. Two explanations are possible. Either the clay fraction consists of a kaolin-like mineral with excess of sesquioxides, or, in some cases, silica; or the minerals present form a mixture of hydrated alumino-silicates and ferro-silicates of varying composition, mixed in some cases with excess of sesquioxides or silica.

Robinson and Holmes (*loc. cit.*), reviewing the results of their analyses of the clay "colloid" from a range of American soils, examined the hypothesis that the primary product of chemical weathering was a mixture of a hydrated

aluminium silicate, $\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot x\text{H}_2\text{O}$ of the kaolinite type, and a hydrated ferric silicate, nontronite, $\text{Fe}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot x\text{H}_2\text{O}$. They found it necessary in the majority of cases to assume the presence of an excess either of silica or sesquioxides.

G. W. Robinson,¹⁶ from a study of the composition of the clay fraction of soils in Wales, concluded that, for soils of primary weathering, the data were not incompatible with the theory that the first product of weathering is a mixture of hydrated silicates of the general formula, $\text{R}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot x\text{H}_2\text{O}$. Variations from the $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio of 2.0 demanded by this hypothesis could be ascribed to differentiation by eluviation or, in the case of soils derived from unconsolidated sediments, concomitant precipitation of silicic acid from river waters. Thus, whilst the average molecular $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio for the clay fraction of soils directly formed by primary weathering was 1.85, the corresponding ratio for the clay fraction of soils derived from unconsolidated aqueous sediments was 2.67.

It should be noted further that, although silica and sesquioxides account for most of the clay fraction, there are also present alkaline and alkaline earth bases. These are present partly as reactive or exchangeable cations and partly as interior non-exchangeable cations. Their significance will be appreciated at a later point in this account.

POSSIBLE MODES OF FORMATION OF CLAY.

In considering the origin of the clay complex by the chemical weathering of silicate minerals, it is evident that two different types of product are possible. On the one hand, the secondary products which constitute the clay complex may originate from silicate minerals by the removal of certain elements or groups of elements from their crystalline lattices, the removal being accompanied in some cases by replacement with hydrogen or hydroxyl groups. The products may retain the original lattice form or, more probably, form different structures.

On the other hand, the origin of the clay complex may consist, as supposed by Van Bemmelen and others, in precipitation from the soluble products of silicate hydrolysis. Whilst, according to the precipitation theory, as originally propounded, the products are absorption compounds of indefinite composition, it is equally possible that the precipitates may be definite, though complex, in composition. The difficulty in assigning formulæ to the clay complex may be due either to the presence of varying proportions of relatively simple crystalline compounds, or to the complex being built up as a highly complicated lattice whose structure cannot be expressed by any simple formula.

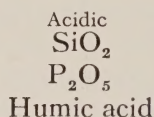
It is probable that both residual products and precipitation products occur in the complex of ordinary soils, but there are, at present, no methods available for their exact differentiation. It is possible that residual products predominate in the weathering complex of strongly leached soils, whilst, with incomplete or impeded leaching, precipitation products predominate. The precipitation products may include, in addition to constituents deposited from solution at the seat of weathering, illuvial material transported from other horizons of the soil or from a ground-water table. These possibilities have been set out on page 40, where the subject is discussed from another aspect.

ANALOGY WITH ISOELECTRIC PRECIPITATES.

Considerable light has been thrown on the constitution of the clay complex by the distinguished investigations of S. Mattson¹⁷, who has studied the precipitates formed under different conditions from solutions analogous to those which might be expected to produce the colloidal complex in soils.

In Mattson's experiments, standard solutions were mixed in varying proportions and the flocculation was observed immediately, and after standing overnight. In addition, the cataphoretic transport under a definite potential gradient was determined ultramicroscopically. Precipitates showing no cataphoretic transport are termed isoelectric and correspond with maximum precipitation.

Data have been obtained for the formation of precipitates in which the basic and acidic components are as follows :—



By carrying out experiments in the presence of hydrochloric acid or sodium hydroxide, the effect of reaction on isoelectric precipitation was also studied.

The following results were obtained in a typical experiment.

TABLE VII.—FLOCCULATION OF ALUMINIUM CHLORIDE AND SODIUM SILICATE SOLUTIONS (MATTSON).

c.c. Solution A	c.c. Solution B	Flocculation		Transport velocity $\mu/\text{sec.}$ at 1 volt/cm.	pH
		After mixing	Overnight		
20	17.4	opalescent	O	...	...
20	17.6	slow	XXXX*	+1.38	5.7
20	17.8	instant	XXXX	+0.76	6.1
20	18.0	instant	XXXX	-0.67	6.4
20	18.5	slow	XX	...	...

A = 5 millimoles AlCl_3 per litre. B = 7.5 millimoles Na_2SiO_3 per litre.

Molecular ratio in isoelectric mixture :— $\text{SiO}_2/\text{Al}_2\text{O}_3 = 2.68$.

pH at isoelectric point :—6.25.

Composition of isoelectric precipitate :— $\text{SiO}_2/\text{Al}_2\text{O}_3 = 1.63$.

*Complete flocculation indicated by XXXX.

The precipitate with 20 c.c. of Solution A and 17.8 c.c. of Solution B is electropositive, as is shown by its direction of transport, whilst the precipitate with 20 c.c. of Solution A and 18.0 c.c. of Solution B is electronegative. The isoelectric precipitate, *ex hypothesi*, shows no cataphoretic movement. Precipitation is most complete at the isoelectric point; but it should be noted, from the discrepancy between the molecular ratio in the isoelectric mixture and in the isoelectric precipitate, that precipitation is incomplete. In the above experiment an excess of silicic acid would remain in solution.

Mattson's results for the different systems obtainable from alumina and ferric oxide on the one hand, and silica,

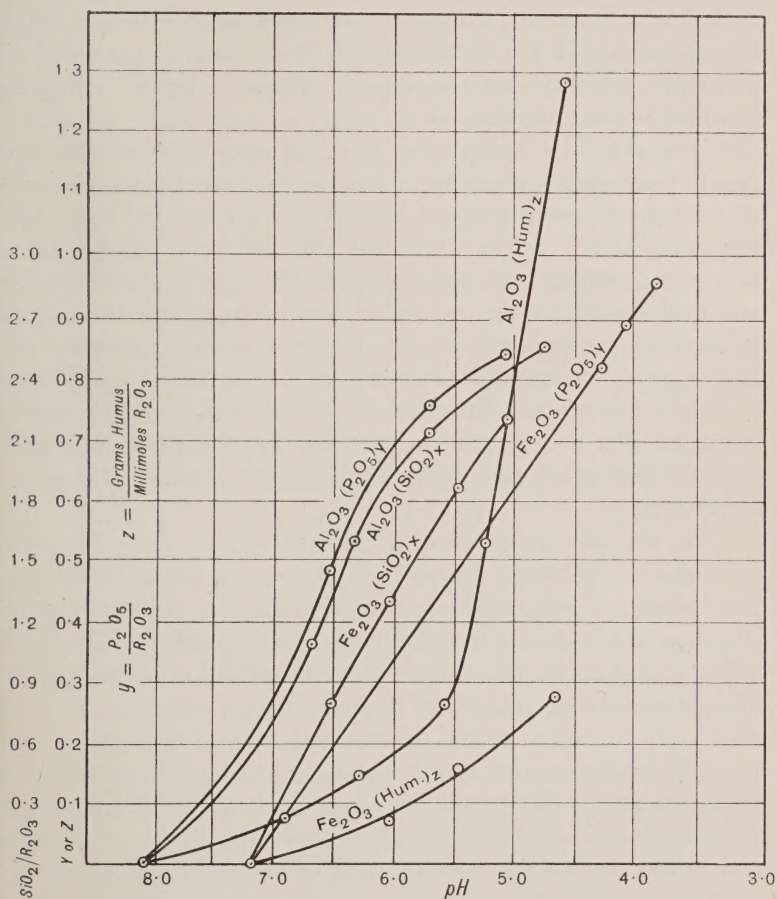


FIG. 3.—Relation between pH and isoelectric composition of aluminium and ferric "phosphates," "silicates," and "humates." (S. MATTSON.)
(From "Soil Science"—by permission.)

phosphoric oxide, and humus on the other, are shown diagrammatically in Fig. 3.

There is, for every pH, a corresponding isoelectric precipitate. It will be seen that for the $SiO_2-Al_2O_3$ system,

the molecular ratio in the isoelectric precipitates varies from about 2.4 at pH 4.5 to zero at pH 8.0. In the case of the $\text{SiO}_2\text{-Fe}_2\text{O}_3$ system, the range of molecular ratios is from about 2.1 at pH 5.0 to zero at pH 7.0. In the presence of bivalent cations—for example, calcium—more siliceous complexes are obtained.

The colloidal complex of the soil may be considered to result from mutual precipitation, at or near isoelectric conditions, from electropositive (basic) sols on the one hand and electronegative (acidic) sols on the other. Whilst it is scarcely possible that sols of opposite sign could originate simultaneously as the result of weathering, and then form isoelectric precipitates as in laboratory experiments, the isoelectric precipitate theory may serve as a useful hypothesis to explain a process in which the intermediate steps postulated by this theory are suppressed, for the isoelectric precipitate represents the most stable product under the given conditions.

If we can assume with Mattson that the inorganic colloidal complex in the soil is analogous to an isoelectric precipitate, a reasonable explanation of the variation in its composition in different soil types is forthcoming. The high silica-sesquioxide ratios observed in the colloidal complexes of arid soils and unleached limestone soils are explained by the high calcium status under such conditions. Under humid conditions the soil is subjected to leaching, whereby the protecting bases are removed and the conditions are represented by the relationships shown in *Fig. 3*. Where conditions favour the development of an acid type of organic matter, as in cool and temperate humid regions—or even, under certain circumstances, in the tropics—the most stable complex is one relatively rich in silica. On the other hand, under tropical conditions, the high temperature causes a rapid mineralisation of organic residues and, although loss of bases occurs, the reaction is more nearly neutral, with the result that a complex rich in sesquioxides is produced.

Mattson's results suggest that phosphoric oxide and humic acids may form integral constituents of the colloidal clay. It is, however, not certain that the whole of the humic material of soils is thus built up with silicic acid and sesquioxides. In highly organic soils, humus may occur independently, though in intimate admixture with the clay complex. It is also possible that sesquioxides may enter into the constitution of the humus complex.

CRYSTALLINE CHARACTER OF CLAY COMPLEX.

Earlier writers on soils, in discussing the nature of the clay constituent, have generally assumed that it has the character of an amorphous gel, but recent work has brought into question this view of the character of clay. For example, E. B. Powell¹⁸ questioned the emulsoid character of colloidal clay on the ground of the smoothness of the curves connecting the viscosity of clay suspensions with temperature.

Further work has shown that, as in the case of many other gels, the constituent material is principally micro-crystalline. The X-ray studies of C. S. Ross¹⁹ revealed the crystalline structure of certain clays and also permitted the identification of some definite minerals as common constituents of clays. The clays examined by Ross fall into two groups, namely, (1) the kaolin group, including kaolinite, $\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 2\text{H}_2\text{O}$ and anauxite, $\text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2 \cdot 2\text{H}_2\text{O}$; and (2) the montmorillonite group, including montmorillonite, $(\text{Ca}, \text{MgO}) \cdot \text{Al}_2\text{O}_3 \cdot 5\text{SiO}_2 \cdot 5\text{H}_2\text{O} \pm$, beidellite, $\text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2 \cdot 0.4\text{H}_2\text{O} \pm$, nontronite, $^* \text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2 \cdot 0.4\text{H}_2\text{O} \pm$, and crystalline halloysite, $\text{Al}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 0.3\text{H}_2\text{O}$. The dominant soil- and clay-forming mineral was considered by Ross to be beidellite, together with its ferric isomorph nontronite. These minerals exhibit a platy crystalline form analogous to mica, and it may be maintained that this structure, involving large surface development and relatively large contact areas between particles, is sufficient to explain the characteristic

*There is some ambiguity as to the formula for nontronite. Robinson and Holmes give the formula as $\text{Fe}_2\text{O}_3 \cdot 2\text{SiO}_2 \cdot 2\text{H}_2\text{O}$.

physical properties of clays. The crystalline nature of clay particles is not inconsistent with their being built up into a gel structure having the property of imbibing water. The structural elements of many other gels are also known to be crystalline.

S. B. Hendricks and W. H. Fry²⁰ have applied the methods of X-ray examination to the colloidal material isolated from a wide range of soils and clays. The crystalline character of the materials examined was shown by comparison with the diffraction patterns obtained from known clay minerals, and it was possible to characterise each of the samples examined. In some cases, notable quantities of quartz were shown to be present, whilst a few samples gave kaolinite patterns. In many cases, montmorillonite, $\text{Mg}(\text{Ca})\text{O} \cdot \text{Al}_2\text{O}_3 \cdot 5\text{SiO}_2 \cdot n\text{H}_2\text{O}$ or beidellite, $\text{Al}_2\text{O}_3 \cdot 3\text{SiO}_2 \cdot n\text{H}_2\text{O}$ were indicated; but since these minerals give identical diffraction patterns, it was not possible to distinguish them. Bentonite, a mineral similar to montmorillonite, was present in certain samples, and halloysite, a mineral similar in composition to kaolinite, in others. Hendricks and Fry consider that there are three main types of clay, namely, (1) the montmorillonite type; (2) the halloysite type; and (3) the Ordovician bentonite type. Colloids with high iron content frequently gave the halloysite pattern. It is presumed that in each type there is some isomorphous replacement of aluminium by iron. Bauxite was found in one sample, whilst it was impossible to identify the minerals of a lateritic clay from Cuba consisting mainly of sesquioxides.

One point which emerges from the work of Hendricks and Fry is that the primary soil minerals, such as feldspars and micas, are not found among the crystalline material of the colloidal fraction, a fact which indicates that the lower limit of their physical sub-division lies above the limit used (0.001 mm.) in defining the colloidal material for this work.

W. P. Kelley, W. H. Dore, and S. M. Brown²¹ compared potash bentonites with colloidal material from certain soils and found similar X-ray patterns.

The results of the examination of soil colloidal material by the X-ray method do not exclude the possibility that the clay complex may contain some amorphous material, in addition to the crystalline minerals identified by this method. It seems possible, nevertheless, that soils and clays may be characterised by the crystalline minerals which they contain, as revealed by X-ray analysis. A systematic survey of the clay minerals from different types of soil will doubtless enlarge the list of minerals and will assist in the more accurate specification of soil characters.

It is to be regretted that most of the investigations by X-ray methods hitherto conducted have been on soils of secondary weathering. Results from soils formed by direct weathering of crystalline rocks would be of the highest interest and significance in elucidating the nature of the process of clay formation.

C. E. Marshall²² has adduced evidence for the crystalline character of clays from a study of their double-refraction properties, and has shown that the bases associated with clays are not, as was supposed by some workers, present at the surface alone, but occupy definite positions in the crystal lattice. From ultra-microscopical observations, Marshall has concluded that, for a given clay, each cationic combination has a characteristic degree of dispersion.

The results described above lead us to regard the clay constituents of soils as composed of a limited number of crystalline minerals of definite chemical composition. That the actual chemical composition of the clay fraction of soils rarely permits the assignment of a definite formula may be attributed to the varying proportions in which the clay minerals occur, and possibly also to the presence of adventitious silica, alumina, and ferric oxide. One important result of these investigations is the proof that a hydrated ferric silicate—possibly nontronite—is present as a clay mineral, and confers on soils in which it occurs a yellowish-green colour. Earlier workers have too readily assumed that the essential ingredients of clay were of an alumino-silicic

character and that iron only occurred as adventitious hydrated ferric oxide. It has been shown also that kaolinite, once considered the characteristic ingredient of clays, is of rare occurrence in clays of epigene origin, though present among the products of pneumatolytic action.

Kelley, Dore, and Brown consider that a complex silicate of the montmorillonite type represents the first product of the chemical weathering of silicates, and that the minerals with lower silica content represent successive stages of degradation consequent on loss of base status. This is in harmony with the views of Mattson as to the effect of bivalent cations on the composition of isoelectric precipitates.

S. Mattson²³ has carried out certain experiments which show that clay complexes low in silica contain their excess of sesquioxides as adventitious constituents. Colloidal clay from a number of red or yellow soils with low silica-sesquioxide ratios was treated with hot saturated aluminium chloride solution. In each case a grey residue having a silica-sesquioxide ratio approximating to 2.0 was obtained. This is the ratio demanded by halloysite, or its ferric isomorph, which has been shown to be a characteristic mineral of highly weathered humid soils.

F. Hardy,²⁴ using a method based on alizarin absorption for the identification and estimation of free sesquioxides in soils, has found that, even in soils which contain a notable excess of sesquioxides as shown by their analyses, the proportions of free sesquioxides are generally small. For example, a lateritic clay from Cuba contained only 7.5% alumina and 8.3% ferric oxide, whilst red soils from Bermuda contained 11.2% alumina and 5.1% ferric oxide. Further, a considerable proportion of the free sesquioxides occurs in the coarser fractions of the soil. Hardy's views appear incompatible with the theory that the weathering complex of sesquioxidic soils consists of a mineral of the general composition $R_2O_3 \cdot 2SiO_2 \cdot xH_2O$ with all the excess sesquioxides present as such, and seems to imply the occurrence of compounds with less than two molecules of silica to each molecule of

sesquioxide. It is, of course, possible that the sesquioxides in excess of the $R_2O_3 \cdot 2SiO_2$ formula may be lightly combined and extractable by the hot aluminium chloride solution used in Mattson's experiments.

UNIVALENT AND BIVALENT CATIONS IN THE CLAY COMPLEX.

We shall discuss, in a subsequent chapter, those reactions of the colloidal complex due to the presence of certain associated univalent and bivalent cations—the so-called exchangeable bases. Our view of the constitution of the clay complex, however, will be incomplete without some reference to the part played by these bases in the structure of the colloidal clay particle.

The exchangeable bases of which we are cognisant in ordinary exchange reactions occur on the surface of the ultimate particles of the colloidal complex. It is certain, however, that univalent and bivalent cations are also present as inner components of the lattice elements.

We have already alluded to the evidence from double refraction adduced by C. E. Marshall (*loc. cit.*) for the occurrence of such components within the lattice of clay minerals. W. P. Kelley, W. H. Dore and S. M. Brown²¹ have demonstrated the presence of inner non-exchangeable cations by experiments in which certain materials were subjected to intensive grinding in a ball mill. The trend of their results will be seen from the following example.

A bentonite from Nevada was found, after replacement of its exchangeable bases by calcium, to contain 106 mg. equivalents of exchangeable calcium per cent. and no other exchangeable bases. After grinding in a ball mill for 72 hours the exchangeable calcium remained unchanged, but there were also present 222 mg. equivalents of exchangeable magnesium per cent., which must have been present originally as inner non-exchangeable lattice cations. Similar results were obtained with other colloidal clays, previously saturated with calcium. In every case, whilst the original

colloidal clay contained calcium as the principal exchangeable cation, the effect of grinding was to mobilise a further quantity of bases, principally magnesium, but including also potassium and sodium. The results of these experiments on soil colloids are shown in Table VIII.

TABLE VIII.—EFFECT OF GRINDING SOIL COLLOIDS (KELLEY, DORE, AND BROWN).

COLLOID	Milligram equivalents replaceable bases per cent.				
	Ca	Mg	K	Na	Total
Clay Adobe, Cal. ...	58.1	6.1	0.3	1.1	65.6
Clay Adobe, Cal. ground 30 hrs. ...	60.4	109.3	29.0	5.4	204.9
Clay loam, Cal. ...	54.7	2.5	0.0	0.0	57.2
Clay loam, Cal. ground 30 hrs. ...	59.9	71.2	25.3	0.0	156.4
Glacial drift soil, Ind.	54.0	2.7	1.0	1.4	59.2
Glacial drift soil, Ind. ground 72 hrs. ...	57.3	21.0	19.3	13.6	111.2
Limestone soil, Tenn.	32.8	2.3	0.0	1.0	36.1
Limestone soil, Tenn. ground 30 hrs. ...	41.7	27.3	26.0	0.0	95.0
Cecil Clay Subsoil, Ala.	14.2	0.0	0.2	3.9	18.3
Cecil Clay Subsoil, Ala. ground 30 hrs. ...	14.0	6.8	7.3	6.0	34.1

From these results, it appears that the preponderance of calcium among the bases found in association with the clay complex by ordinary replacement methods by no means reflects the true character of the complex, but rather the character of the superficial cations. Within the lattice, the predominant cation appears to be magnesium, and it is suggested that the actual mineral involved in these reactions is montmorillonite, $\text{Mg}(\text{Ca})\text{O} \cdot \text{Al}_2\text{O}_3 \cdot 5\text{SiO}_2 \cdot n\text{H}_2\text{O}$ or the similar mineral bentonite, in both of which some isomorphous replacement of aluminium by iron may occur. H. W. Kerr²⁵ had previously suggested that the constituent responsible for exchange reactions has the composition

$\text{H}_2\text{O}.\text{Al}_2\text{O}_3.6\text{SiO}_2.8\text{H}_2\text{O}$, the hydrogen of the first H_2O group being replaceable.

GENERAL REVIEW OF THE CONSTITUTION OF THE CLAY COMPLEX.

From the above account of the development of ideas on the clay complex, it is evident that the indefinite absorption compound hypothesis which replaced the old kaolinite hypothesis must now give place to a conception of the clay complex as consisting of a few minerals of definite crystalline structure. With them may be associated, in certain cases, free hydrated sesquioxides. The characteristic clay minerals are secondary in character, since they do not occur in crystalline rocks, but are distinct from the zeolites, to which, however, they show certain resemblances. They may all be represented by the general formula $(\text{R}_2\text{O}_3)_x(\text{SiO}_2)_y(\text{H}_2\text{O})_z$. Whilst certain of the clay minerals, such as halloysite and beidellite, appear to be non-reactive, in minerals of the montmorillonite and bentonite type, univalent and bivalent cations are constituents of the lattice and, where they occur at the surface, are capable of exchange reactions with the cations of electrolytic solutions. The exchangeable cations occurring at the surface are not necessarily present in the proportions in which they occur within the mineral lattice. Whilst there are indications that the predominant cation within the lattice is magnesium, calcium is usually the predominant external (exchangeable) cation.

The clay minerals include alumino-silicates and ferri- or ferro-silicates. In addition, minerals may occur in which there is an isomorphous replacement of aluminium by iron.

The information at present available on the clay minerals is only of a preliminary character, and one of the principal tasks in the immediate future is the survey of the minerals of the principal world groups and sub-groups of soils. In the meantime, the existing information on the composition of the clay complex of different soils permits certain broad distinctions to be drawn.

The broadest distinction is between clays with high silica-sesquioxide ratios and clays with moderate or low silica-sesquioxide ratios. Any attempt to indicate a critical limit is necessarily provisional, but there is some justification for considering a molecular silica-sesquioxide ratio of 2.0 as significant. Soils whose clay fractions have silica-sesquioxide ratios greater than 2.0 are generally greyish or brownish-grey in colour in the absence of organic matter, and probably do not contain appreciable proportions of sesquioxides, apart from illuvial accessions. On the other hand, soils whose clay fractions have ratios less than 2.0 generally betoken by their brownish, yellowish, or reddish mineral colours the presence of free ferric hydroxide, and, by inference, aluminium hydroxide. Such soils only show grey colours under conditions favouring reductive processes, as in ground-water soils—where, however, drainage impedance more usually results in a siliceous type of clay.

We may, therefore, conceive a clay of the first class as consisting of a range of minerals in which the montmorillonite-bentonite type is the more predominant the more marked is its siliceous character. On the other hand, a clay of the second class may be considered to be made up of one or more minerals of the type of halloysite or its ferric isomorph, together with adventitious sesquioxides.

Clays of the first class may be developed under a number of different conditions.

- I. They may be formed by primary weathering from crystalline rocks under conditions which hinder the removal of silicic acid. This may occur under the following circumstances :—
 - (a) Leaching may be restricted owing to an excess of evaporation over rainfall in semi-arid and arid climates.
 - (b) Drainage may be impeded, as in ground-water soils, or in soils developed from parent materials with impervious strata, such as many of the boulder clays of Northern Europe.

- II. They may be formed by the precipitation of dissolved silicic acid concomitantly with suspended matter in estuarine, lacustrine, and marine sediments. The clay of immature soils derived from calcareous sediments appears, from the writer's experience, to be of an exceptionally siliceous type. It will be remembered that Mattson has demonstrated the siliceous character of the isoelectric precipitates formed in the presence of bivalent ions.
- III. They may be formed through the removal from the clay complex, by leaching, of associated sesquioxides. This can take place in the following processes:—
- (a) Podsolisation under acid humus.
 - (b) Solotisation, consequent on the hydrolysis of the sodium clay of alkaline soils and the removal of sodium hydroxide by leaching, whereby the clay complex becomes unstable, and loses sesquioxides. In this process, there may be an actual liberation of silica.

In clays of this class, it may be supposed that minerals of the siliceous montmorillonite-bentonite type are present as characteristic, if not dominant, ingredients. A complete survey of the different types of siliceous clays is, however, necessary before this can be safely assumed. It would be of particular interest to ascertain if the removal of sesquioxides in podsolisation and solotisation leads to the formation of new minerals.

Clays of the second class are formed under humid conditions, in which the lowering of the base status by leaching leads to instability of the more siliceous complexes, and may even result in the partial or complete degradation or decomposition of the $R_2O_3 \cdot 2SiO_2 \cdot xH_2O$ complex into free silicic acid and sesquioxides. These conditions obtain in all humid climates in situations with free drainage, apart from those regions where peaty humus is developed and podsolisation occurs.

Concurrently with degradation, silicic acid may be removed under climatic conditions which favour rapid mineralisation of plant residues and a slightly acid soil reaction. According to the intensity of the removal, this gives rise to brown earths, red and yellow earths, and lateritic soils. This may be contrasted with the process under climatic conditions which favour the formation of acid humus from the residues of vegetation. There, the leaching is more strongly acid and sesquioxides are removed as humus-protected soils.

The red and yellow soils of tropical and sub-tropical climates probably differ from the brown earths of temperate climates both in the extent to which free sesquioxides have been liberated and also in their degree of hydration.

It has been suggested that the reactive constituent of clays is a siliceous mineral of the montmorillonite-bentonite type. Although clays of the second class are less reactive than clays of the first class, such reactivity as they possess implies the presence of constituents containing exchangeable bivalent and univalent cations. The constituents are probably less siliceous than the minerals of the montmorillonite class, but still remain to be investigated.

We may further distinguish a type of clay which has been enriched by illuvial accumulation. The B horizons of podsol profiles and the upper horizons of lateritic profiles are demonstrably of this type; but it is also possible that illuvial deposition of sesquioxides may occur to some extent in tropical red loams, red earths, and, if we accept the views of A. Reifenberg²⁶, in terra rossa.

Certain tropical desert soils, in spite of the restriction of leaching by arid conditions, may show bright red colours suggesting the liberation of sesquioxides. It seems possible that in such cases the stability of the siliceous clay minerals is affected by the high temperature. The writer has found silica-sesquioxide ratios considerably greater than 2.0 in certain sediments of the Trias, a formation which is considered to have been laid down under desert conditions.

SIGNIFICANCE OF SILICA-ALUMINA AND SILICA-SESQUIOXIDE RATIOS.

During recent years much attention has been given to the composition of the clay fraction as a means of characterising soils. The most significant information is given by the molecular silica-alumina ($\text{SiO}_2/\text{Al}_2\text{O}_3$) or silica-sesquioxide ($\text{SiO}_2/\text{Al}_2\text{O}_3 + \text{Fe}_2\text{O}_3$) ratio. Whilst some workers use the silica-alumina ratio, others prefer the silica-sesquioxide ratio. The writer is strongly of opinion that the latter ratio is preferable. The use of the silica-alumina ratio implies that the ferric oxide, known to be present in the clay fraction, is uncombined with silica, and present as adventitious ferric oxide. Such an assumption cannot be made, and we are bound to assume that the clay minerals include not only alumino-silicates but also ferri-silicates, and possibly ferro-silicates.

The silica-sesquioxide ratio, though of great value in distinguishing different types of clay, must not be assumed to give a complete specification of the material to which it relates. Two clays may have the same silica-sesquioxide ratio and even the same alumina-ferric oxide ratio and yet be markedly different in constitution. Complete definition can only be attained when it is possible to record all the minerals and the relative proportions of each present.

In general, the silica-sesquioxide ratio of clays of humid tropical climates tends to be low, indicating the tendency to laterisation, whilst that of clays of arid regions and of cold regions tends to be high.

E. M. Crowther,²⁷ from a review of the data of W. O. Robinson and R. S. Holmes (*loc. cit.*), relating to the composition of the colloidal clay of representative American soils, has attempted to assess the relative effect of humidity and temperature on the silica-alumina ratio. There appears to be a negative correlation of this ratio with rainfall, and a positive correlation with temperature. But no general quantitative relationship is likely to be discoverable, even although the enquiry be restricted to soils of primary weathering,

unless it is possible to allow for the effect of age and the possibility of erosion. In humid tropical soils, the tendency is for soils to become more sesquioxidic in character with age; but it is difficult to see how this tendency can be given quantitative expression and introduced into a general formula. The complications introduced by erosion, whereby illuvial horizons are laid bare at the surface, are equally obstructive in the way of obtaining a general relationship between silica-alumina or silica-sesquioxide ratio, humidity, and temperature.

One climatic effect on the silica-sesquioxide ratio of clay may be, however, a fruitful subject for investigation, and that is the tendency of eluviation in modifying the composition of the clay complex. Whilst under cool humid conditions the tendency is towards an impoverishment of the upper horizons in sesquioxides, resulting in a clay complex relatively rich in silica, under humid tropical conditions, the losses by eluviation fall on the silicic acid and the clay complex tends to be enriched in sesquioxides.

Abundant examples of the decrease in silica-sesquioxide ratio from the surface downwards are obtainable from profile examinations in humid temperate and cool regions. Recently, L. D. Baver and G. D. Scarseth²⁸ have studied the composition of colloidal clay from profiles in Alabama. Their results for soils derived from an unconsolidated parent material with a high silica-sesquioxide ratio plainly show an increase in the ratio from the surface downwards, betokening an eluviation of silicic acid. The northern limit of this type of weathering is placed at the 61°F. mean annual isotherm. Between this and the region of podsollic eluviation lies a transition zone, that of the brown forest soils, in which there is no marked predominance of silicic acid or sesquioxides in the losses by eluviation, and in which, consequently, no marked variations in the silica-sesquioxide ratio of the clay complex are to be expected in passing down the profile.

BIBLIOGRAPHY.

- ¹ JOSEPH, A. F., "Clays as soil colloids." *Soil Sci.*, 1925, **20**, pp. 89-94.
- ² HENDRICKS, S. B., and FRY, W. H., "The results of X-ray and microscopical examination of soil colloids." *Soil Sci.*, 1930, **29**, pp. 457-476.
- ³ BRADFIELD, R., "The chemical nature of colloidal clay." *J. Amer. Soc. Agronom.*, 1925, **17**, pp. 253-270.
- ⁴ ATTERBERG, A., "Die Plastizität und Bindigkeit liefernden Bestandteile der Tone." *Int. Mitt. Bodenk.*, 1913, **3**, pp. 291-330.
- ⁵ ROBINSON, G. W., "The form of mechanical composition curves of soils and other granular substances." *J. Agric. Sci.*, 1924, **14**, p. 626.
- ⁶ VAN BEMMELEN, J. M., "Die Absorptionsverbindungen und das Absorptionsvermögen der Ackererde." *Landw. Vers. Stat.*, 1888, **35**, pp. 67-136.
- ⁷ STREMMER, H., *Monatsber. Deutsch. Geolog. Gesell.*, 1910, pp. 122-128.
- ⁸ HARRASSOWITZ, H., "Laterit, Material und Versuch erdgeschichtlicher Auswertung." *Fortschr. Geol. Paleontol.*, 1926, **4**, pp. 253-566.
- ⁹ GANSSEN, R., "Die charakterisierung des Bodens nach der molekularen Zusammensetzung des durch Salzsäure zersetzlichen silikatischen Anteiles des Bodens." *Int. Mitt. Bodenk.*, 1913, **3**, pp. 529-571.
- ¹⁰ ROBINSON, W. O., and HOLMES, R. S., "The chemical composition of soil colloids." *U.S. Dept. Agric. Bull.*, 1924, No. 1311.
- ¹¹ HALL, A. D., "The mechanical analysis of soils and the composition of the fractions resulting therefrom." *J. Chem. Soc.*, 1904, **85**, pp. 950-963.
- ¹² HENDRICK, J., and OGG, W. G., "The composition of the soil and of the mineral particles which compose it." *J. Agric. Sci.*, 1916, **7**, pp. 458-469.
- ¹³ MARCHAND, B. DE C., and VAN DER MERWE, C. R., "On the composition of the fractions separated by mechanical analysis from some Transvaal soils." *S. Afr. J. Sci.*, 1925, **22**, pp. 104-118.
- ¹⁴ ROBINSON, G. W., Unpublished data.
- ¹⁵ EDGINTON, G., Analysis in M. S. ANDERSON, "The influence of substituted cations on the properties of soil colloids." *J. Agric. Res.*, 1929, **38**, p. 567.
- ¹⁶ ROBINSON, G. W., "The nature of clay and its significance in the weathering cycle." *Nature*, 1928, **121**, p. 903.
- ¹⁷ MATTSON, S., "The laws of colloidal behavior." IV.—Isoelectric precipitates. *Soil Sci.*, 1931, **31**, pp. 57-77.
- ¹⁸ POWELL, E. B., "Soil colloids as simple suspensions." *Soil Sci.*, 1928, Vol. IV, pp. 555-561.
- ¹⁹ ROSS, C. S., "The mineralogy of clays." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 555-561.
- ²⁰ HENDRICKS, S. B., and FRY, W. H., "The results of X-ray and mineralogical examination of soil colloids." *Soil Sci.*, 1930, **29**, pp. 457-476.

- ²¹ KELLEY, W. P., DORE, W. H., and BROWN, S. M., "The nature of the base-exchange material of bentonite soils, and zeolites as revealed by chemical investigations and X-ray analysis." *Soil Sci.*, 1931, **31**, pp. 25-56.
- ²² MARSHALL, C. E., "The orientation of anisotropic particles in an electric field." Part I.—General. Part II.—Application to the determination of the double refraction of clays. *Trans. Faraday Soc.*, 1930, **26**, pp. 173-189.
- ²³ MATTSON, S., "The laws of colloidal behavior." V.—Ion adsorption and exchange. *Soil Sci.*, 1931, **31**, pp. 311-331.
- ²⁴ HARDY, F., Studies in tropical soils. I. Identification and approximate estimation of sesquioxide components by adsorption of alizarin. *J. Agric. Soc.*, 1931, **21**, pp. 150-156.
- ²⁵ KERR, H. W., "The identification and composition of the soil alumino-silicate active in base exchange soil acidity." *Soil Sci.*, 1928, **26**, pp. 385-398.
- ²⁶ REIFENBERG, A., "Entschung der Mediterran-Roterde." *Kolloid Beihefte*, 1929, **28**, pp. 1-93.
- ²⁷ CROWTHER, E. M., "The relationship of climatic and geological factors to the composition of the clay and the distribution of soil types." *Proc. Roy. Soc. B.*, 1930, **107**, pp. 10-30.
- ²⁸ BAVER, L. D., and SCARSETH, G. D., "Suptropical weathering in Alabama as evidenced in the Susquehanna fine sands." *Soil Res.*, 1931, **2**, pp. 288-307.

CHAPTER V.

BASE EXCHANGE AND OTHER REACTIONS OF THE COLLOIDAL COMPLEX.

ABSORPTION.

THE colloidal complex is the seat of the most important chemical reactions of the soil, for, except in the case of calcareous, gypseous, and saline soils, the non-colloidal material is relatively inert, and chiefly significant from the physical standpoint.

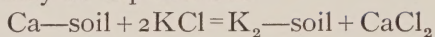
One of the most characteristic properties of colloids is their ability to absorb dissolved substances from solution. This property is shown markedly by soils. It is, however, important to distinguish two types of absorption. The first type, more properly termed adsorption, depends on the change in interfacial tension produced by the presence of a dissolved substance in the liquid phase. Substances which lower interfacial tension tend to accumulate at an interface, whilst substances which raise interfacial tension show negative absorption. This type of absorption by the soil, though demonstrable, is of negligible moment in the pedogenic processes.

The second type of absorption, on the other hand, is of the highest importance in the study of the soil, and is due to reactions which are essentially chemical and ionic in character, but located at the surface of colloidal particles. The enormous development of specific surface implied by colloidal subdivision gives to these reactions their distinctive character. As usually understood, they are comprised under the term *base exchange*, but it should be realised that these

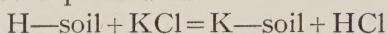
are not the only manifestations of surface chemical activity of which the soil colloidal complex is capable.

BASE EXCHANGE.

When an ordinary soil is allowed to react with the solution of a neutral salt, such as potassium chloride, a proportion of the potassium is absorbed or fixed by the soil and an equivalent amount of calcium is displaced into solution. The reaction may be expressed as



Certain soils, known as acid soils, under similar circumstances, give up acid hydrogen to the solution, and the reaction may be expressed as



These and related phenomena are described as base exchange reactions and have afforded a great body of information as to the nature of the reactive constituents of soils. The cations which are concerned in base exchange are termed *exchangeable* or *replaceable bases*.

The modern study of the subject dates from the work of H. S. Thompson¹ and J. T. Way² in the middle of the nineteenth century. The stimulus to their work was provided by the fear that the newly introduced soluble fertilisers might be washed out of the soil in drainage. Thompson first showed that ammonium sulphate in solution undergoes a reaction with the soil whereby ammonium is fixed and calcium liberated. The careful and exhaustive investigations of Way demonstrated the chemical equivalence of absorption and displacement when soils react with solutions of neutral salts, and located the seat of these reactions in what we now know as the colloidal or absorbing complex. It was also shown by Way that the speed of absorption was very great and that, at a certain concentration of the dissolved salt, an upper limit of absorption was reached.

The absorptive properties of soils have been extensively studied since the discoveries of Way and Thompson, and many theories have been advanced to account for the observed

facts. Liebig maintained that absorption by soils was a purely physical process analogous to the absorption of dyes by charcoal. Whilst this type of absorption, depending on surface tension effects, may undoubtedly occur in soils, Liebig's theory did not provide a satisfactory explanation of the reactions of soils with soluble salts. Way considered these reactions as essentially chemical, but the development of colloidal chemistry during the latter half of the last century proved an important help to the study of absorption reactions, and suggested the explanation that these phenomena, though chemical, are distinguished from ordinary chemical reactions by the fact that they are localised at surfaces.

We may now consider some of the characteristics of the reactions between soils and solutions of electrolytes. When a soil is allowed to react with a solution of an electrolyte such as potassium chloride, potassium is absorbed by the soil and an equivalent amount of other ions, principally calcium, is displaced into the solution. The chloride ion is not only unabsorbed, but may actually show negative absorption, whereby the chloride concentration in the equilibrium solution is greater than in the gel water associated with the soil colloids. Absorption is, however, incomplete; for, even with the most dilute solutions, a certain proportion of the absorbable ion remains in solution. G. Wiegner,³ from a review of available data, showed that the relationship between the concentration of an ion in the solid phase—the soil—and the concentration in the equilibrium solution could be expressed by the so-called Freundlich equation: $y = KC^{1/p}$, where y is the amount in unit weight of soil, including that originally present, C the concentration of the ion in the solution, and K and p are constants.

Regarding, now, the ions liberated from the soil by a displacing ion, these will also be distributed between soil and solution, and displacement will be only partial. It is possible, however, by successive treatments of a quantity of soil with a salt solution, removing the equilibrium solution

after treatment, to obtain an almost complete displacement of the exchangeable bases originally present by the cation of the electrolyte used. This can be most readily effected by continuous leaching of the soil, as in the method of D. J. Hissink⁴ for the determination of exchangeable bases. The exchangeable bases in 25 g. of soil can be removed by leaching on a filter with one litre of normal sodium chloride solution. At the conclusion of this process, the exchangeable bases originally present in the soil have been replaced by an equivalent amount of sodium, and the soil may then be termed a *sodium soil*.

Hissink and others have shown that base exchange is practically instantaneous — a circumstance which is readily intelligible on the assumption that it is a surface reaction.

In Table IX data are given for the content of exchangeable bases of some typical soils.

TABLE IX. — EXCHANGEABLE BASES IN MILLIGRAM EQUIVALENTS* IN TYPICAL SOILS.

SOILS	Ca	Mg	K	Na	Authority
25 Dutch clay soils ...	30.0	5.0	0.8	2.5	D. J. Hissink (<i>loc. cit.</i>)
17 Scottish soils ...	9.95	0.776	0.242	0.269	A. M. Smith ⁵
7 Neutral soils, U.S.A.	13.92	4.83	0.75	1.48	W. P. Kelley and S. M. Brown ⁶
5 Alkali soils, U S A.	0.0	0.80	1.65	6.88	W. P. Kelley and S. M. Brown ⁶
6 Acid soils, U S A.	1.06	0.68	0.13	0.51	W. P. Kelley and S. M. Brown ⁶

It will be seen that, apart from alkaline soils, the principal base is calcium, accompanied by smaller amounts of magnesium, potassium, and sodium. Hissink found that, cal-

*In the statement of results of exchangeable base determination it is convenient to use milligram equivalents rather than the actual percentages of exchangeable bases. A milligram equivalent is the equivalent weight in milligrams, so that 1 mg. equivalent of CaO = 0.028 g. A percentage of CaO is thus converted to mg. equivalents by dividing by 0.028. The mg. equivalent content of the other bases is computed in the same way.

culating all bases to milligram equivalents, the average composition of the exchangeable bases of 25 Dutch clay soils was

Ca	Mg	K	Na
79	13	8	3

Different figures are obtained for other types of soil, but, in normal soils, the four bases occur in proportions not widely different from those found by Hissink. In saline and alkaline soils, on the other hand, sodium is the predominant exchangeable base.

BASE-SATURATION AND UNSATURATION.

Soils which have long been in equilibrium with calcium carbonate are said to be saturated with exchangeable bases. Under field conditions, the maximum capacity of a soil for exchangeable bases is the content of total bases when the soil is in equilibrium with excess of calcium carbonate. Any base present in excess of this amount appears in the soil in the form of carbonate.

Where calcium carbonate is absent, the content of exchangeable bases may fall below that representing saturation and the soil is said to be base-unsaturated or, simply, unsaturated. Such a condition may be produced under laboratory conditions by leaching a soil with dilute acids, such as hydrochloric or acetic acid, whereby the exchangeable bases go into solution as chlorides or acetates, and the soil becomes completely desaturated of bases. A similar process occurs in nature where soils, in the absence of free calcium carbonate, are subjected to the leaching action of percolating waters. The solvent action in such cases is augmented by the presence of dissolved carbon dioxide, which has originated from the respiration of plant roots and from the microbiological decomposition of plant residues in the soil. In extreme cases, as for example, in very humid climates and with highly pervious soils, an almost complete removal of exchangeable bases, resulting in desaturation, may take place. For example, in the uplands of Wales, uncultivated soils fre-

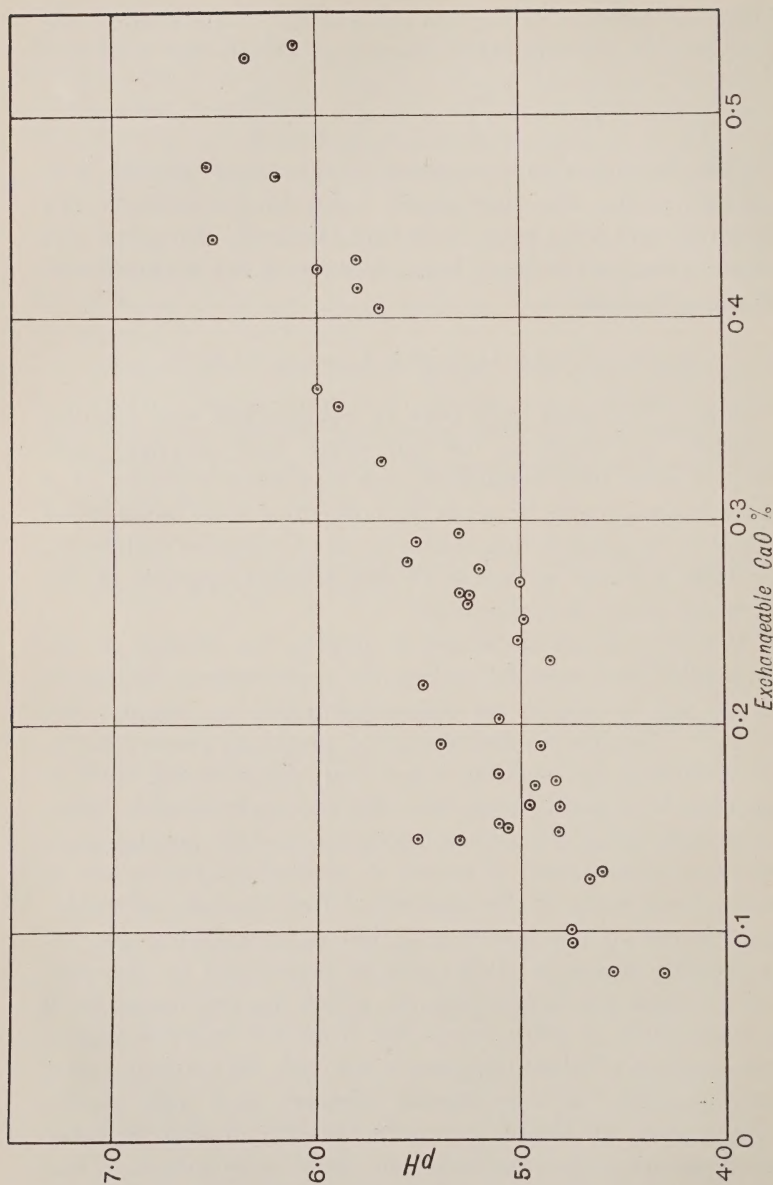


FIG. 4.—Relationship between pH and exchangeable CaO of Anglesey soils.

quently contain less than 0.05% of exchangeable calcium oxide. Figures as low as 0.01% may be encountered in light sandy soils.

Unsaturation may sometimes exist in soils which contain free calcium carbonate, notably in soils derived from hard limestones. The calcium carbonate may occur in relatively coarse hard fragments, and saturation is only attained in the immediate vicinity of such fragments, whilst the body of the soil is subjected to loss of exchangeable bases by leaching. Similarly, the remains of former dressings of lime or chalk may persist as coarse fragments, although the soil as a whole has become unsaturated.

EXCHANGEABLE BASE CONTENT AND SOIL REACTION.

The greater the degree of unsaturation of a soil, the more marked are its acid properties. A convenient measure of the acidity of soil is the reaction of its aqueous suspension, expressed as its pH., i.e., the negative logarithm of its hydrogen-ion concentration. As the exchangeable base content of a given soil falls, its reaction will become increasingly acid, that is to say, its pH will progressively fall.

Since calcium accounts for about 80% of the reaction value of the exchangeable bases in soils, a close relationship should exist between the reaction, as shown by the pH value, and the exchangeable calcium, in soils with the same type of colloidal complex present in approximately the same proportions. This relationship is seen in *Fig. 4*, which shows graphically the relationship between pH and exchangeable calcium oxide for a number of soils of the Anglesey series. These soils contain about 17-20% of clay and about 5 to 8% of organic matter. The $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio of the clay is about 1.9.

It will be seen that there is a general tendency to a linear relationship. The "scatter" of the experimental points may be attributed to the varying proportions of organic matter and clay present and to variations in the character of the clay itself.

Although, among soils of similar composition, there is an obvious relationship between the reaction of the soil, as expressed by its pH, and its content of exchangeable calcium oxide, a general expression for this relationship is not easily obtained, since the acidic properties of a soil vary not only with the relative proportions of mineral and organic matter in the colloidal complex, but also with the composition of the mineral absorbing complex. Whilst the pH of base-unsaturated organic matter may be 4.0 or even less, that of the so-called clay acid is higher and depends ultimately on the relative proportions of acid (SiO_2 , P_2O_5 , humic acid) and basic (Al_2O_3 , Fe_2O_3) groups present. Clay colloids rich in silica, such as those of bentonite and certain alluvial clays will be more acid when desaturated than clay colloids rich in sesquioxides, such as those of laterites and tropical red earths. (See also p. 121.)

DEGREE OF UNSATURATION.

Much ingenuity has been expended on devising methods for determining the extent of unsaturation of acid soils, i.e., the amount of base required to produce saturation. The difficulty consists in arriving at a satisfactory definition of saturation. For, even when a soil contains such a proportion of exchangeable bases that its reaction is neutral ($\text{pH}=7.0$), it can still combine with more bases. The case is parallel with that of a weak acid such as acetic acid. If a strong alkali such as soda be added to a given amount of acetic acid in dilute solution, the quantity of soda added when the neutral point is reached is less than the equivalent of acetic acid present. When the soda is exactly equivalent to the acetic acid the reaction of the solution is markedly alkaline owing to hydrolytic dissociation.

D. J. Hissink⁷ has sought to measure the degree of unsaturation of soils by determining the amount of base with which a soil can combine in the presence of such an excess of alkali that hydrolytic dissociation is suppressed. Under ordinary conditions, however, a soil would never arrive at

such a state of base-saturation, and it has therefore been proposed to determine the amount of base required to bring a soil to the point of neutrality. This can readily be ascertained by electrometric titration, by adding a standard solution of alkali to a soil suspension, and determining the pH potentiometrically after each addition.

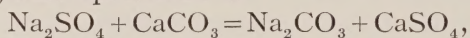
In another class of methods, soil is allowed to react with calcium carbonate. The carbonate decomposed, measured by the amount of carbon dioxide evolved, is a measure of the degree of unsaturation and, if equilibrium is reached, actually represents, when the base equivalent is added to the equivalent of the bases originally present, the *exchange capacity*, i.e., the base content of the soil in the highest state of saturation attainable under ordinary conditions.

The question of base-unsaturation is of considerable importance from the economic standpoint, since, by means of it, an estimate is sought of the amount of lime or calcium carbonate necessary as a dressing in order to bring the soil to an optimum base status. The difficulty of the problem lies in discovering what is the optimum base status, or the minimum base status consistent with fertility, since this varies according to climate and cropping. Whilst for certain crops, such as potatoes, a base status definitely below saturation, corresponding with a markedly acid reaction, may be sufficient for satisfactory growth, other crops, such as lucerne, appear to require a higher base status. The actual relationship between pH and plant growth has been studied by many workers. Whilst it may be generally stated that plant growth is adversely affected by very acid or very alkaline reactions ($\text{pH} < 5.0$ or > 8.5), individual crops show wide and ill-defined optima. Actually, the relationship is not simply between plant growth and soil reaction, for calcium supply and the possible effect of toxic aluminium in acid soils enter into the problem. From the practical viewpoint, then, it is not yet certain whether the important factor is the soil reaction, its content of exchangeable bases, or its deficit from base saturation.

SODIUM SOILS.

In the foregoing account, allusion has been made to the occurrence of soils containing sodium as the dominant exchangeable base. It will now be convenient to examine the chemistry of such soils. We shall consider, at a later point, their general properties and structure under field conditions.

It has long been known that the presence of sodium salts in the soil is attended, under certain conditions, by marked changes in physical properties. These were attributed to the formation of sodium carbonate. The reaction, as conceived by Hilgard, was represented as



assuming sodium sulphate to be the salt originally present. This theory of the formation of the sodium carbonate known to occur in "alkali" soils proved obviously inadequate, since it failed to explain why a reaction which proceeds to a negligible extent under laboratory conditions should result in the production of notable amounts of sodium carbonate under field conditions.

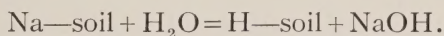
The mechanism of the reaction has been satisfactorily explained by the investigations of A. De Domenicis⁸, K. Gedroiz⁹, A. A. J. De Sigmond¹⁰, and A. B. Cummins and W. P. Kelley,¹¹ and will be clear from the following discussion.

If a quantity of a base-saturated soil containing calcium as the principal exchangeable base be treated with a normal solution of sodium chloride, calcium will be displaced from the soil and an equivalent amount of sodium will be absorbed. By prolonged leaching, it is possible to displace all the original bases present by sodium, and the soil may now be described as a sodium soil. It should be added that, in the case of acid or unsaturated soils, complete replacement of hydrogen is not readily effected. Only by a very prolonged leaching could an acid soil be changed into a sodium soil.

If the treatment of the calcium soil with normal sodium chloride solution be carried out by leaching on a filter, the

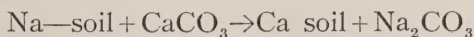
rate of percolation is not sensibly changed by the transformation to the sodium soil so long as there is an excess of sodium chloride present. In other words, the normal sodium chloride maintains the sodium soil in the flocculated condition. On removing the excess of sodium by washing, however, it is seen that a marked change takes place. The rate of filtration becomes considerably slower, whilst the filtrate becomes turbid and, if an appreciable amount of organic matter be present, dark coloured. There is also a change in reaction, for, whilst the leachings with normal sodium chloride are approximately neutral, the dark-coloured turbid runnings, obtained after the excess of sodium chloride has been removed, are markedly alkaline to phenolphthalein.

These phenomena, which are of the greatest importance for the study of alkaline soils, have been differently explained by different workers. Essentially, however, the process is a case of salt hydrolysis. The sodium soil is a salt of a weak acid and a strong base. In the presence of an excess of sodium chloride, its tendency to hydrolyse is suppressed. When the concentration of the sodium chloride is sufficiently reduced, a partial hydrolysis takes place which may be thus expressed :



The alkaline reaction, due to the liberation of small quantities of sodium hydroxide, favours the dispersion or deflocculation of the soil, and the solution of organic matter in the form of dark coloured sodium humate.

Under natural conditions, the removal of excess of sodium salts from a sodium soil is accompanied by the formation of sodium carbonate. This may occur either by the reaction :



or else by the reaction of carbon dioxide with the sodium hydroxide liberated by hydrolysis. The sodium carbonate is not formed by the direct reaction of a soluble salt with calcium carbonate.

ENERGY OF ABSORPTION OF DIFFERENT CATIONS.

The energy of absorption of cations by the colloidal complex varies considerably. The order of absorption of the principal ions is $H > Ca > Mg > K > Na$. The replacement of exchangeable sodium by exchangeable hydrogen is, thus, more rapid than the replacement of exchangeable hydrogen by exchangeable sodium. The relative absorption of calcium and sodium is illustrated by certain data obtained in the Sudan Government Laboratories¹². Sodium-saturated clay was repeatedly treated with solutions containing calcium and sodium chlorides which were decinormal with respect to chloride, but contained varying equivalent ratios of calcium to sodium. The ratios of calcium to sodium in the clay were determined after each treatment. The following is a selection from the results obtained :—

Ca/Na in solution	2.33	1.50	1.00	0.8	0.42	0.20	0.11	0.058
Ca/Na in clay	14.2	10.6	9.3	6.8	4.9	2.66	1.59	0.76

It is seen that the clay in equilibrium with the solution containing equal equivalent amounts of calcium and sodium contained 9.3 times as much calcium as sodium, and that in order to obtain an excess of sodium in the soil the ratio of sodium to calcium in the percolating solution had to be raised to the order of 15 : 1.

The relative energy with which bases enter the absorbing complex is also shown by some results obtained by A. F. Joseph and H. B. Oakley¹³, who allowed equimolecular solutions of calcium, potassium, and sodium chlorides, of strength 0.5 normal with respect to chloride to percolate through a soil. With calcium and potassium chlorides, the ratio Ca/K attained in the soil was 0.9; with calcium and sodium chlorides the Ca/Na ratio was 6; and with potassium and sodium chlorides the K/Na ratio was 6. It appears, therefore, that the energy of replacement of calcium is slightly greater than that of potassium and that both greatly exceed sodium in this respect.

The mutual displacement of cations associated with colloidal particles is governed by their degree of hydration.

This is illustrated by data obtained by H. Jenny¹⁴, who determined the exchange constants of a number of ions against calcium permutite and obtained the following values for K :—

Li	Na	K	NH ₄	Rb	Cs	H
3.500	6.066	9.896	11.425	11.550	13.112	58.117

The corresponding constants for bivalent ions were:—

Mg	Ca	Sr	Ba
29.802	48.128	49.882	69.137

Hydrogen shows the greatest energy of replacement, and this renders it exceedingly difficult to determine the content of exchangeable hydrogen in acid soils by leaching with salt solutions. Nevertheless, a certain amount of displacement occurs, and this gives rise to what is sometimes termed *exchange acidity*. Methods for determining exchange acidity have been proposed; but it should be realised that they do not give the total content of exchangeable hydrogen. Estimates of the degree of unsaturation of a soil obtained by measuring the pH of a suspension in 0.1 N-potassium chloride are equally uninformative as to the actual base-acid status and must be considered as having a purely empirical value.

CONTRIBUTION OF CLAY AND ORGANIC MATTER TO EXCHANGE CAPACITY.

It has been already stated that the property of base exchange in soils resides in the colloidal complex, which is composed of clay and organic matter. The relative contributions of these two categories of soil colloidal material to exchange capacity is a matter of considerable interest and different estimates have been made.

Varying figures are obtained according to the nature and origin of the materials, and the methods employed for determining exchange capacity. For colloidal clay, the figures obtained by S. Mattson¹⁵ range from 0.164 to 1.102 mg. equivalents per gram of colloidal clay, the lowest figure

being given by a highly sesquioxidic clay and the highest figure by bentonite, a highly siliceous clay. The estimates for the exchange capacity of humus vary from about 2.5 to 4.5 mg. equivalents per gram.

Considering saturation to be represented by the base-status of soils which have long been in equilibrium with excess of calcium carbonate, R. Williams¹⁶ has sought to determine the relative base-binding capacity of clay and organic matter by comparing the total exchangeable base content of a number of calcareous soils with their content of clay and organic matter. The data are shown in Table X.

TABLE X.—RELATION BETWEEN TOTAL EXCHANGEABLE BASES, CLAY CONTENT, AND ORGANIC CARBON CONTENT FOR CARBONATE SOILS.

SOIL	Clay %	Organic Carbon		CaCO ₃ %	Total bases in mg. equivalents calculated by		Total bases in mg. equivalents %
		Total	Oxidis- able		Formula 1	Formula 2	
Vaynol Clay	63.5	0.64	0.14	13.6	39.1	37.08	36.80
B 1459 ...	54.0	1.54	0.96	8.3	37.79	36.83	36.67
G. 109 ...	41.7	4.45	2.43	7.0	44.02	39.08	40.00
Harpenden	24.7	1.06	0.59	4.1	18.80	17.80	18.58
G. 26 A ...	39.3	2.54	2.06	3.6	33.96	35.39	34.45
G. 115 A ...	27.5	1.94	1.45	3.3	24.51	24.82	25.89
G. 130 A ...	34.3	3.14	2.22	3.1	33.85	33.53	33.03
Llysfasí ...	27.0	2.29	1.83	3.1	25.81	26.92	23.60
Rendzina	59.6	1.63	1.30	2.7	41.39	42.16	46.40
Groes Ffordd	9.3	1.89	1.58	2.6	13.90	15.26	16.07
F. 85 ...	24.8	2.09	1.36	2.4	23.55	22.71	22.50
G. 137 ...	21.2	2.39	1.91	2.2	22.96	24.21	23.20
D. 129 ..	33.6	3.34	2.62	1.7	34.34	35.65	32.86
G. 206 ...	3.6	7.21	3.07	8.0	46.25	32.79	35.72
G. 178 A ...	47.1	3.62	2.95	3.3	43.32	45.43	38.67
G. 57 ...	22.9	2.98	1.84	0.8	26.61	24.64	13.57
F. III ...	19.0	3.28	2.20	0.5	25.80	24.69	19.29
Ab 3-5 ...	22.4	3.17	2.36	0.2	27.19	27.64	14.46

Organic carbon has been used as a measure of the humus and two figures have been obtained in each case, namely, total carbon, and organic carbon removable by prolonged boiling with 4% hydrogen peroxide. The latter determination was made in order to avoid the error in total organic carbon introduced, in some cases, by the presence of the fragments of coal and cinder which are so frequent in cultivated soils.

By the use of a graphical method, equations were obtained for the relationship between total bases (T), clay (K), total carbon (C_t), and oxidisable carbon (C_o). The relationships thus found are :

$$T = 0.57 K + 4.55 C_t \dots\dots\dots (1)$$

$$\text{and } T = 0.57 K + 6.3 C_o \dots\dots\dots (2)$$

The calculated values are shown in columns 6 and 7 respectively.

It will be seen that the agreement between the calculated values for total bases and those actually found is fairly satisfactory for the first thirteen soils. G. 206 contains a considerable proportion of coal fragments, which leads to a high figure being given by Formula 1. On the other hand, the figure by Formula 2 is rather low.

The last four soils all show experimental values markedly below calculated values. In the case of G. 178 A, most of the carbonate is present as coarse hard fragments, whilst in the case of the remaining three soils the small amounts of carbonate present probably represent fragments surviving from calcareous dressings. The result is that these soils are definitely unsaturated.

The exchange capacity indicated for clay, namely, 0.57 mg. equivalents per gram falls within the limits found by S. Mattson (*loc. cit.*), and is in good agreement with a figure obtained by E. M. Crowther and J. K. Basu¹⁷ for Woburn soils*.

Assuming soil organic matter to contain 58% of carbon, the exchange capacities are 2.64 mg. equivalents per gram of

*Since the above results were obtained, analyses have been made of the clay fractions of the soils in question. Considering only those soils in which base-saturation may be assumed, it appears that the SiO_2/R_2O_3 ratios form a well spaced range from 1.69 in the case of D. 129 to 4.78 in the case of G. 26A. Mattson, using a method depending on leaching to equilibrium with calcium chloride, found a strong correlation between SiO_2/R_2O_3 ratio and exchange capacity. The present results show no evidence of such a correlation and suggest that the maximum base-binding capacity of clay under field conditions is not markedly affected by its composition. If this be generally true, it should be permissible to use one of the formulæ given above for obtaining the field saturation capacity of a soil, and hence, by comparison with its content of exchangeable bases, its degree of saturation.

total organic matter and 3.65 mg. equivalents per gram of oxidisable organic matter. Both figures fall within the limits given at the beginning of this section.

EXCHANGEABLE BASES AND THE STABILITY OF THE CLAY COMPLEX.

The nature and content of exchangeable bases in a soil have an important bearing on the stability of the colloidal matter.

Dealing with the colloidal clay complex, four main cases may be distinguished.

(1) The soil exhibits a high base status. Here the clay complex possesses complete stability and there is no differentiation of the soil into horizons of varying silica-sesquioxide ratio. This case is seen in the tshernosems and chestnut-coloured earths of semi-arid regions and those carbonate soils of humid regions which contain an excess of calcium carbonate in intimate admixture.

(2) The soil has a low base status and, in addition, owing to the presence of acid humic material, as in cold humid regions, has a markedly acid reaction. Here the clay complex is unstable and the tendency is for the removal of sesquioxides by percolating waters, with the production of a podsol profile.

(3) The soil has a low base status, but owing to the rapid mineralisation of plant residues, under tropical conditions, the reaction does not become markedly acid. Here the clay complex is also unstable, but the tendency is towards the removal of silicic acid by percolating waters, resulting in the production of tropical red earths, which are characterised by clay fractions relatively rich in sesquioxides.

(4) The soil contains an excess of sodium among its exchangeable bases. The colloidal complex remains stable in the presence of excess of sodium salts and calcium carbonate. In the absence of sodium salts, the sodium clay undergoes hydrolysis with formation of sodium hydroxide and the clay acid which, if no calcium carbonate is present

to re-form the calcium soil, eventually undergoes a change, known as solotisation, analogous to podsolisation.

In the brown earths, it would appear that the base status is sufficiently low to induce an incipient degradation of the clay complex, resulting in a liberation of silicic acid and sesquioxides, but the acidity of the percolating waters is not sufficiently pronounced to lead to any differentiation into horizons. The brown earths are probably the mid-latitude analogues of the tropical red loams and red earths.

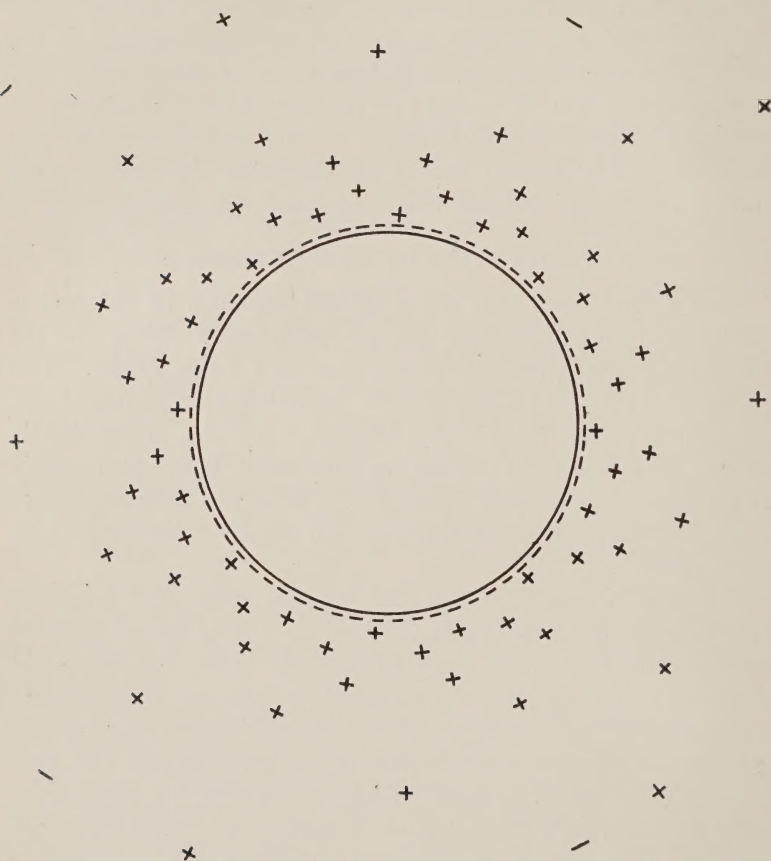
PROPERTIES OF SUSPENSIONS.

We have seen that the colloidal material of the soil is built up from (a) acidoid groups, principally silica, and (b) basoid groups, principally alumina and ferric oxide. The properties of a colloidal particle suspended in a liquid are governed primarily by the character of the outer layer, which, so far as soil colloids are concerned, is predominantly of an acidoid or anionic character. With the acidoid groups are associated corresponding cations, the so-called exchangeable cations.

The structure of a colloidal particle suspended in water is shown diagrammatically in *Fig. 5*. The inner shell of anions and the outer shell of cations form the so-called double layer. Whilst the anionic groups are fixed as the outer shell of the colloidal particle, the corresponding cations arrange themselves with respect to the particle and the surrounding liquid in a distribution depending on their own ionic activity, and their degree of hydration. The distribution is also governed by the presence of electrolytes according to the well known Donnan equilibrium, whereby the product of the concentrations of an ionic pair must be the same in all parts of the system.

The nett result is that the cationic sheath forms a diffuse layer rather than a definite shell of cations all at the same distance from their corresponding anions. Some of the cations will form undissociated combinations with the sheath anions. Indeed, it is probable that only a small proportion

of the exchangeable cations are actually dissociated and capable of contributing to the diffuse double layer. For



+ Cations, including Hydrogen ions

- Anions, including Hydroxyl ions

FIG. 5.—Diagram of a colloidal particle with anions and cations of the double layer.

example, S. Mattson¹⁸ has calculated that, in the case of a colloidal particle of radius $45.5 \text{ m}\mu^*$ with 502,208 exchange-

*₁ $\text{m}\mu = 0.000001 \text{ mm}$.

able cations, a dissociation of only 857 ions would account for the observed electrical potential.

Since a proportion of cations is relatively remote from the particle, there is an excess of negative charges. In other words, the particle is negatively charged. The magnitude of the change will be governed by the character of the colloidal surface and by the extent to which dissociation has taken place. The potential difference (P.D.) between the inner and the outer sheath is given by

$$\text{P.D.} = ed/Dr^2,$$

where e is the charge on the particle, d the mean radial distance between the anionic sheath and the cationic sheath, D the dielectric constant, and r the radius of the particle. For a given particle, since e/Dr^2 is approximately constant, the potential will depend mainly on d , the radial distance between the two sheaths. This is governed by the degree of hydration of the cations present. The more heavily hydrated these are, the greater is the distance between them and the inner layer. The order of hydration of the univalent cations is $\text{Li} > \text{Na} > \text{K} > \text{Rb} > \text{Cs} > \text{H}$, and this is also the order for the potential of the corresponding particles of clay.

It has been stated in the discussion of mechanical analysis that the fractions of a soil are separated by making use of their differing settling velocities. The validity of a mechanical analysis depends on the attainment of the so-called prime particle structure, in which each ultimate particle settles independently. In very fine suspensions, a change can be brought about whereby the separate particles aggregate into larger compound structures, which settle at a greater rate than that of the particles of which they are composed. This change, called coagulation, can be induced by the addition of electrolytes. For example, the addition of a few c.c. of normal calcium chloride to a litre of clay suspension is rapidly followed by the formation of floccules which, in a few minutes, settle to the bottom, leaving a clear supernatant liquid. The stability of suspensions depends on their particles being free to settle independently of each other.

The particles of a suspension carry electrical charges, usually negative, and the condition for independent behaviour is that their potential shall be above a certain critical value. When, for any reason, the potential of the particles falls below this value, the electrostatic repulsion is insufficient to prevent the association of particles into compound aggregates. The nature of the force whereby such aggregates are held together has been differently explained as cohesion, as residual valency, or as linkage through common cations.

Two stages are distinguished in coagulation. The first stage, which is reached when the potential of the suspended particles falls below the critical potential, is slow coagulation. If the potential is further lowered, coagulation becomes increasingly rapid until it reaches a maximum in instantaneous coagulation, or flocculation.

Coagulation of suspensions results from a decrease in the electrokinetic potential of their suspended particles and can be brought about in one of the following ways: (1) By addition of electrolytes to the suspension medium. This results in a repression of the dissociation of the outer cations, whereby cations moving freely in the solution rejoin the colloidal particle, with the consequence that a proportion of the free charges on the inner sheath anions is neutralised. (2) By decreasing the degree of hydration of the cations, as for example, by addition of alcohol. This has the effect of decreasing the radial distance between the inner and the outer sheaths. The same effect is produced when highly hydrated cations such as lithium and sodium are replaced by less hydrated cations such as potassium, calcium, or hydrogen. (3) By increasing the dielectric constant of the medium.

The stability of suspensions in terms of their associated cations is in the order $\text{Li} > \text{Na} > \text{K} > \text{Rb} > \text{Cs} > \text{Mg} > \text{Ca} > \text{Sr} > \text{Ba} > \text{H}$, the well known Hofmeister Series, and this is also the order of hydration of the ions.

Since the stability of suspensions depends on the charge of their particles, it follows that, so far as we are dealing with

negative suspensions, the more marked is the acidoid character of the colloidal particles, the greater, *ceteris paribus*, will be the electrical charge and the higher its electrokinetic potential. This is, in fact, the case: clays having a high silica-sesquioxide ratio are more stable than clays having a low silica-sesquioxide ratio. Thus, whilst it is possible to prepare stable suspensions of highly siliceous clays with hydrogen as the cation, with clays of a lateritic character, i.e., with a low silica-sesquioxide ratio, the hydrogen clays are unstable, and stability can only be attained by the introduction of a highly hydrated ion such as sodium.

CATAPHORESIS.

One of the consequences of the electrical charge possessed by suspended colloidal particles is that they move in an electrical field, the speed of migration being proportional to the electrokinetic potential of the charged particle. Measurements of the speed and direction of cataphoretic movement, made with the aid of the ultra-microscope, have proved of the highest value in the study of the variations in the electrical properties of colloidal suspensions. (Cf. p. 79.)

ELECTRODIALYSIS.

The exchangeable bases of soils may be separated by electrodialysis, a procedure which is essentially similar to the electrolysis of electrolytes. In the method as originally developed by S. Mattson¹⁹, a three-compartment cell was used, the soil suspension being in an inner chamber separated from the outer compartments, containing the electrodes, by parchment membranes. The voltage used was 220 volts (D.C.) and the current was regulated by resistances in order to avoid too great a rise in temperature. The exchangeable cations appear in the cathode chamber, whilst the diffusible anions appear in the anode chamber. After 22 hours' dialysis, a Sharkey clay soil gave 28.937 mg. equivalents of base in the cathode chamber, and 3.375 mg. equivalents of acid in the anode chamber. At the conclusion of dialysis, the base-free hydrogen soil remains.

Many simplified methods of electrodialysis have been proposed, for example, that of M. L. M. Salgado and G. W. Chapman²⁰, in which the soil is contained in a Jena glass filtering crucible through which water is allowed to percolate. The anode is a spiral of gold wire and rests on the soil, whilst the cathode is of copper gauze and is held below the filter disk.

The content of exchangeable bases determined by electrodialysis is generally in good agreement with that determined by displacement methods. It has been objected by some workers that electrodialysis results in a decomposition of the anionic portion of the absorbing complex. M. Trénel²¹ found that permutite gels, bereft of bases by electrodialysis, underwent a certain decomposition into silicic acid and sesquioxides. The desaturation thus attained is therefore not completely reversible.

SUSPENSION EFFECT.

It has been already stated that only a proportion of cations belonging to the outer sheath of suspended colloidal particles is in the dissociated state. Although these ions are held in the vicinity of the particles, their concentration is yet susceptible of measurement by potentiometric methods. The hydrogen-ion concentration of a suspension is thus made up of the hydrogen-ion concentration due to ions in the suspension medium and the hydrogen-ion concentration due to ions held by the suspended particles. The total hydrogen-ion concentration, according to G. Wiegner and H. Pallmann²², may be expressed by the relationship:

$$C^1_H = C^0_H + K\xi,$$

where C^1_H is the actual hydrogen-ion concentration, C^0_H is the hydrogen-ion concentration of the suspension medium, ξ is the percentage concentration relative to the most concentrated suspension used, and K is a constant, meaning the increase in hydrogen-ion concentration for each 1% increase in concentration of the suspension.

We may thus conceive the exchangeable hydrogen of suspended colloidal particles as existing in three categories, namely,

- (1). Hydrogen of undissociated complexes.
- (2). Hydrogen-ions in the actual cationic sheath.
- (3). Hydrogen-ions existing freely in the suspension medium.

It may be presumed that the equilibrium between the three classes of ions is dynamic so long as the particles are in actual suspension. By analogy, it may be supposed that the same three categories hold for other exchangeable cations.

The suspension effect shown by hydrogen-ions is also shown by hydroxyl-ions. The contribution of the sheath-ions to the hydrogen-ion concentration is shown by the difference between the hydrogen-ion concentration of the suspension and that of the medium from which the particles have been removed by sedimentation or centrifuging. If alkali is added to an acid suspension, the suspension effect becomes progressively less until, in the region of the neutral point, the pH of the suspension is equal to that of the suspension medium. With further addition of alkali, an alkaline suspension effect is observed, i.e., the suspension has a more alkaline reaction than the medium. The particles now have an excess of hydroxyl-ions, which are, of course, located in the inner layer.

AMPHOTERIC CHARACTER OF THE COLLOIDAL COMPLEX.

We have hitherto considered those reactions of the colloidal complex in which its behaviour is that of a weak insoluble acid, or acidoid, as it has been termed. The work of S. Mattson²³ has shown that this is only one aspect of its reactions and that it possesses the character of an ampholyte. This is shown by its ability to absorb anions and to pass over from an electronegative to an electropositive colloid.

This behaviour is illustrated by a series of experiments in which equal weights of electrodialysed soil colloids were treated with solutions of ammonium chloride, ammonium

sulphate, and ammonium phosphate, respectively, the proportion of cation to anion being varied by adding different amounts of ammonia and the respective acid.

Some typical data are given in Tables XI and XII. The data in the first table refer to a lateritic clay having a molecular $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio of 0.31, whilst those in the second table refer to a siliceous clay having a molecular $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio of 3.18.

TABLE XI.—ABSORPTION OF NH_4 AND Cl IONS BY NIPE COLLOID, 3% SUSPENSION (MATTSON).

Mg. equivalents added		Mg. equivalents absorbed		Cataphoresis $\mu/\text{sec.}$ at 1 volt/cm.	pH
NH_4	Cl	NH_4	Cl		
2.242	2.00	0.039	0.00	—1.30	7.25
2.121	2.00	0.021	0.003	—0.30	6.7
2.00	2.00	0.014	0.011	+0.047	6.1
2.00	2.10	0.012	0.024	+0.82	5.8
2.00	2.20	0.006	0.038	+1.51	5.3
2.00	2.40	0.007	0.059	+1.78	4.05

TABLE XII.—ABSORPTION OF NH_4 AND Cl IONS BY SHARKEY COLLOID, 2.8% SUSPENSION (MATTSON).

Mg. equivalents added		Mg. equivalents absorbed		Cataphoresis $\mu/\text{sec.}$ at 1 volt/cm.	pH
NH_4	Cl	NH_4	Cl		
4.178	2.00	0.661	0.00	—2.16	6.8
3.089	2.00	0.376	0.00	—1.68	5.6
2.00	2.00	0.111	0.001	—1.01	3.2
2.00	2.20	0.090	0.001	—0.60	3.1
2.00	2.40	0.081	0.001	—0.43	3.0
2.00	2.80	0.066	0.004	—0.25	2.8

In both cases, with decrease in pH, cationic absorption decreases and anionic absorption increases. There is also a

decrease in the electronegative character of both colloids as shown by their cataphoretic velocity; but whilst, in the case of the Nipe colloid, there is a transition through an isoelectric point to an electropositive suspension, in the case of the Sharkey colloid, although the negative cataphoresis decreases, no isoelectric point is indicated. Similar results were obtained for the absorption of SO_4 and PO_4 ions, the energy of absorption being in the order $\text{Cl} < \text{SO}_4 < \text{PO}_4$.

In general, colloids having low silica-sesquioxide ratios (i.e., acidoid-basoid ratios) more readily absorb anions and pass over to the electropositive state than colloids having high silica-sesquioxide ratios.

The ultimate pH of a soil, i.e., the pH after all diffusible anions and cations have been removed, as in electrodialysis, depends on the relative dissociation of the acidoid and basoid components. Where $K_{\text{acidoid}} > K_{\text{basoid}}$, the reaction is on the acid side of neutrality. This is the case even with the most markedly basic soil colloids such as that of the Nipe clay. Mattson (*loc. cit.*) obtained the following pH values for suspensions of four typical colloids after electrodialysis.

Nipe	Cecil	Sassafras	Sharkey	Humus
6.65	4.75	4.63	3.73	3.72

The corresponding $\text{SiO}_2/\text{R}_2\text{O}_3$ ratios for the four mineral colloids were 0.31, 1.34, 1.89, and 3.18 respectively.

We have seen that Mattson considers the soil colloids to be analogous to isoelectric precipitates. The colloidal material of the soil, according to this hypothesis, has a complex constitution and varies in properties according to the proportion of acidoid groups. The principal acidoid groups are silicic acid and humic acids, and the principal basoid groups, alumina and ferric oxide.

The variation in properties with differences in the relative proportions of acidoid and basoid groups is now well established. But there appears to be a definite conflict between Mattson's conception of a colloidal complex of indefinite structure and composition and the conception, which

has emerged from the investigations by X-ray methods, of the crystalline structure of soil colloids, according to which the clay complex is pictured as a mixture of identifiable crystalline minerals, some possessing and some devoid of base exchange properties. Mattson himself reconciles the two conceptions by postulating an indefiniteness of composition in the surface layer of particles, although admitting that the law of definite proportions may hold within the lattice.

A highly siliceous mineral of the montmorillonite type has been indicated as the carrier of the base exchange properties of soil. Does this imply that the variations in acidic character with $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio are due to variations in the proportions of the montmorillonite mineral to the less reactive clay minerals? Does a mineral of the montmorillonite type occur in soils with low $\text{SiO}_2/\text{R}_2\text{O}_3$ ratios or are there minerals of the kaolinite type with base exchange properties? These are problems which await solution in the future, and for which the views of Mattson may furnish a provisional working hypothesis.

BIBLIOGRAPHY.

- ¹ THOMPSON, H. S., "On the absorbent power of soils." *J. Roy. Agric. Soc.*, 1850, **11**, pp. 68-74.
- ² WAY, J. T., "On the power of soils to absorb manure." *J. Roy. Agric. Soc.*, 1850, **11**, pp. 313-379; *ibid.*, 1852, **13**, pp. 123-143.
- ³ WIEGNER, G., "Zum Basenaustausch in der Ackererde." *J. Landw.*, 1912, **60**, pp. 110-157, pp. 197-222.
- ⁴ HISSINK, D. J., "Beitrag zur Kenntnis der Adsorptionsvorgänge im Boden." *Int. Mitt. Bodenk.*, 1912, **12**, pp. 81-172.
- ⁵ SMITH, A. M., "The exchangeable bases in some Scottish soils." *J. Agric. Sci.*, 1925, **15**, pp. 466-475.
- ⁶ KELLEY, W. P., and BROWN, S. M., "Replaceable bases in soils." *Univ. of Calif. Technical Paper* 15, 1925, pp. 1-39.
- ⁷ HISSINK, D. J., "Base exchange in soils." *Trans. Faraday Soc.*, 1925, pp. 551-617.
- ⁸ DE DOMENICIS, A., "Terreni salsi e terreni alcalini." *Staz. Sperim. Agr. Ital.* 1918, **41**, pp. 103-161.
- ⁹ GEDROIZ, K., "Die Kolloid-chemie in Fragen der Bodenkunde." II.—Die Bildung der Soda im Boden. *Russ. Journ. Exper. Landw.* 1912, **13**, pp. 363-420.
- ¹⁰ DE SIGMOND, A. A. J., "Contribution to the theory of the origin of alkali soils." *Soil Science*, 1926, **21**, pp. 455-479.

- ¹¹ CUMMINS, A. B., and KELLEY, W. P., "The formation of sodium carbonate in soils." *Univ. of Calif. Techn. Paper* 3, 1923, pp. 1-35.
- ¹² *Report of Government Chemist, Sudan*, 1930, p. 13.
- ¹³ JOSEPH, A. F., and OAKLEY, H. B., "The properties of heavy alkaline soils." *J. Agric. Sci.*, 1929, **19**, pp. 121-131.
- ¹⁴ JENNY, H., *Kolloid. Beih.*, 1927, **23**, p. 428.
- ¹⁵ MATTSON, S., "The relation between the electrokinetic behavior and the base exchange capacity of soil colloids." *J. Amer. Soc. Agronomy*, 1926, **18**, pp. 458-470.
- ¹⁶ WILLIAMS, R., Unpublished results.
- ¹⁷ CROWTHER, E. M., and BASU, J. K., "Studies in soil reaction." VIII.—The influence of fertilisers and lime in the replaceable bases of a light acid soil after fifty years' continuous cropping with barley and wheat. *J. Agric. Sci.*, 1931, **21**, pp. 689-715.
- ¹⁸ MATTSON, S., "Laws of colloidal behavior." II.—Cataphoresis, flocculation and dispersion. *Soil Sci.*, 1929, **28**, pp. 373-409.
- ¹⁹ MATTSON, S., "Electrodialysis of the colloidal soil material and the exchangeable bases." *J. Agric. Res.*, 1926, **33**, pp. 553-567.
- ²⁰ SALGADO, M. L. M., and CHAPMAN, G. W., "The determination of exchangeable bases by electrodialysis." *Soil Sci.*, 1931, **32**, pp. 199-216.
- ²¹ TRÉNEL, M., "Elektrodialyse und das Problem der mineralische Bodenazidität." *Trans. Int. Soc. Soil Sci.*, Comm. II, 1929, Vol. B., pp. 144-153.
- ²² WIEGNER, G., and PALLMANN, H., "Über Wasserstoff- and Hydroxylschwärmionen von suspendierte Teilchen und dispergierte Ultramikronen." *Z. Pflanz. Düng.*, 1930, **16A**, pp. 1-57.
- ²³ MATTSON, S., "Laws of colloidal behavior." VI.—Amphoteric behavior. *Soil Sci.*, 1931, **32**, pp. 343-365.

CHAPTER VI.

SOIL ORGANIC MATTER.

ORIGIN FROM PLANT RESIDUES.

IN our general view of the constitution of the soil, we have discussed the relationship of organic matter to the other constituents present. We must now consider its constitution in connexion with those changes effected by microbiological agencies, whereby plant residues become incorporated with the soil as humus.

The proportion of organic matter in the soil varies considerably. Whilst in the case of peat soils, the dry matter may be almost entirely organic, in ordinary cultivated soils, organic matter rarely amounts to more than 15% of the total dry weight. The highest figures are obtained in grassland soils under humid climates, and the lowest figures under deficient rainfall in hot climates. The variations in organic matter content in the soil profile are also considerable. There is a general decrease in organic matter from the surface downwards in all soils, except humus podsols with horizons of humus accumulation. Whilst the decrease in organic matter with depth is generally gradual, in some cases there is an abrupt and almost discontinuous decrease at a certain point in the profile.

The relationship of the organic matter content of the soil to such factors as climate, soil texture, water conditions, and cultivation will be more appropriately discussed when the mechanism of the decomposition of plant residues has been examined.

The humus of soils arises mainly from plant residues. There is, of course, a certain contribution from animal remains and excreta, notably in the case of cultivated soils.

Under forest vegetation, the principal source of the soil organic matter is leaf-fall. In certain types of forest, however, undergrowth and ground vegetation make a considerable contribution from their residues. Indeed, it is probably only in the case of dense coniferous forest that leaf-fall forms the sole source of organic matter. Beech forests which, during the summer months, may be devoid of ground vegetation, may yet have an appreciable ground flora in spring.

Under steppe or prairie conditions, the soil is occupied by gramineous and herbaceous plants, whose growing season is limited by drought or winter cold. The dead remains of each season's growth become added to the soil in the form of leaves, stems, and roots. Artificial grassland, in this respect, can be classed with steppe and prairie. Under cultivation, organic matter is added to the soil in the form of stubble, trash, and dressings such as farmyard or pen manure.

Plant residues added to the soil differ in certain respects from fresh plant materials. They represent, in the main, structures which have fulfilled their functions and are generally almost devoid of starches, sugars, and fats; for these have already been stored in seeds or tubers, or have been transformed into fibrous materials. The principal constituents of plant residues, then, are old structural elements, together with small amounts of protein, traces of waxes and resins, and ash constituents.

Oat straw, which is typical of all plant residues contains, according to H. Pringsheim¹:—

Celluloses	Pentosans	Lignins	Crude Protein	Waxes	Ash
35.43	21.33	20.40	4.70	2.02	4.81

The elemental composition of some plant residues is shown by the following data given by Dvůrák². They relate to dry matter in each case.

	C.	H.	N.	O.	Ash
Pine needles	41.96	3.98	1.42	21.07	31.57
Oak leaves	49.11	6.12	1.71	29.38	13.68
Wheat Straw ...	47.01	5.66	0.82	38.61	7.90
Lucerne	43.28	5.86	1.95	38.54	10.37

MICRO-ORGANISMS OF THE SOIL.

Whilst purely chemical decompositions play a certain part in the decomposition of plant residues, particularly in tropical soils, these changes are mainly due to the activity of the soil fauna and flora. The more important groups of the soil fauna are :—

- I. Protozoa.
- II. Nematodes.
- III. Worms, insects, etc.

The more important groups of the soil micro-flora are :—

- I. Algæ and Diatoms.
- II. Fungi, including Actinomycetes.
- III. Bacteria.
 - A. Autotrophic, deriving their energy from the oxidation of simple inorganic compounds and their carbon from carbon dioxide. They include :
 - i. Nitrifying bacteria.
 - ii. Sulphofying bacteria.
 - iii. Iron bacteria.
 - B. Heterotrophic, deriving their carbon from complex organic compounds. They include :
 - i. Nitrogen fixing bacteria.
 - a. Symbiotic, e.g., *Bact. radiculicola*.
 - b. Free living.
 - 1. Aerobic, e.g., *Azotobacter*.
 - 2. Anaerobic, e.g., *Clostridium*.
 - ii. Bacteria concerned with the process of ammonification.
 - iii. Cellulose bacteria and other bacteria decomposing fibrous materials.

The numbers of bacteria in soils have been estimated by methods which it is not necessary to describe here. That the figures obtained represent actual conditions in the natural soil may be received with some doubt; but they may serve as a basis for comparison. In ordinary soils from 2 to 200 millions of bacteria per gram of soil have been recorded, the

highest figures being obtained in highly cultivated and heavily manured soils.

Considerable seasonal fluctuations are observed in the bacterial numbers in soils. The highest figures are obtained under the joint effect of moisture and warmth, whilst drought and cold, particularly the latter, exert a depressing effect. There is a rapid decrease in bacterial numbers with depth, and complete sterility is reached within a few feet of the surface.

Fungi are present in considerably smaller numbers than bacteria. They are most abundant in organic soils with acid reaction. In such soils the decompositions are mainly the work of fungi rather than of bacteria.

The work of the lower animals may be considered in the first place. Under British conditions, this type of attack is mainly through the agency of earth worms, who feed on the soil, and, by passing it through their bodies, perform an important task in comminuting and mixing the plant residues with the mineral portion of the soil. It has been calculated by Darwin that, in the course of half a century, the whole of the soil of a pasture down to ten inches has been brought to the surface in the form of casts. It is generally considered that the rôle of earth worms in the decomposition of organic matter in the soil is mainly mechanical.

In soils with markedly acid reaction, earth worms do not occur, and the residues of plants tend to accumulate at the surface as a peaty layer. This is seen in the podzols of coniferous forests and heaths.

In the tropics, the most important invertebrate connected with the organic matter cycle is the termite or white ant.

The changes induced by earth worms, white ants, and similar organisms, although they involve certain chemical decompositions, are principally of importance through their effect in distributing organic matter throughout the soil profile. In the absence of such organisms, as in highly acid soils, the added organic matter remains as a sharply defined

layer resting on a mineral soil which is generally deficient in organic matter.

By reason of their enormous numbers, the organisms constituting the microflora are of chief importance in the decompositions affecting soil organic matter. Before proceeding to an account of their activities it may be helpful to draw certain distinctions between their conditions of development and function.

The first distinction to be drawn is between *autotrophic* and *heterotrophic* organisms. Autotrophic organisms are those which obtain their carbon from carbon dioxide and their energy from the oxidation of inorganic compounds. Examples of these are the nitrifying bacteria.

Heterotrophic organisms, on the other hand, obtain their carbon from complex organic compounds and their energy from the oxidation of such compounds. Included in these are the bacteria which oxidise cellulosic materials, and the bacteria concerned in the fixation of atmospheric nitrogen. Heterotrophic bacteria may be divided into *aerobic* organisms working in the presence, and *anaerobic* organisms, working in the absence or deficiency of oxygen.

The transformations affecting soil carbon and soil nitrogen are the work of autotrophic and heterotrophic, of aerobic and anaerobic organisms. Grossly considered, these changes may be grouped into two classes, namely, oxidative decompositions and humification.

OXIDATIVE DECOMPOSITION OF ORGANIC MATTER.

In oxidative decomposition, organic matter is oxidised to carbon dioxide and water. These changes, which are the work of many organisms, are favoured by good aeration, and are most active between 35 and 40°C. Moisture is necessary, since bacteria live in a fluid medium, but excess of moisture by excluding air from the soil, slows down and may even inhibit bacterial oxidation. The presence of calcium carbonate would appear to favour oxidative decomposition of organic matter; but it is by no means certain that oxida-

tive decomposition cannot proceed equally well in the absence of calcium carbonate, provided there is a sufficient supply of calcium, combined in readily available forms with the clay and organic matter of the soil, to serve as a source of calcium and to maintain a reaction favourable to bacterial life. A high calcium status, by favouring granular structure, improves aeration. Soils containing large amounts of free calcium carbonate are particularly well aerated and oxidation of organic matter proceeds in them with great rapidity, provided adequate moisture be present.

The oxidative decompositions will therefore proceed with the greatest rapidity where optimum moisture conditions, optimum temperature, and a sufficiency of lime in the soil coincide. Under such conditions, organic matter readily disappears, a circumstance which is reflected in the low organic matter content of cultivated soils of hot countries. Even in colder climates, the aeration induced by cultivation can lead to the rapid disappearance of organic matter from the soil. An example of this is given by some data obtained by the writer in North Wales.

A cultivated soil from the uplands of Caernarvonshire showed a loss on ignition of 16.6 per cent. The soil on the other side of the fence, which presumably corresponded with that from which the field had been reclaimed 50 years before, gave an ignition loss of 43.1%. Although ignition loss is an untrustworthy measure of organic matter, the difference between these figures does give an approximate idea of the rapid loss which has taken place, and this in spite of the fact that the cultivated soil had been from time to time under long grass leys and had received the ordinary dressings of farmyard manure during the arable break.

The oxidative decomposition of organic matter results in the liberation in the soil of large amounts of carbon dioxide. In so far as microbiological activity can be correlated with general fertility, the rate of evolution of carbon dioxide may afford important information as to biological conditions in the soil. E. J. Russell³ found a correlation between the

rate of absorption of oxygen and the productivity of soils in the same locality.

The carbon dioxide produced in the soil by the oxidation of plant residues is the most important single source of this compound in nature. Its importance both for the nutrition of plants and also for the processes of soil formation can hardly be over-estimated. Dissolved in the soil moisture or combined in the form of hydrogen-carbonates, it plays an important part in the vertical translocation of materials within the soil profile. In the absence of humic acids from peaty organic matter, carbonic acid is the principal acid concerned in soil formation.

Whilst the oxidative decomposition of organic matter in the soil is mainly the work of micro-organisms, it is probable that purely chemical decompositions also play a part. Such decompositions are likely to be of greatest significance under tropical conditions in soils subjected to periods of such intense drought as to suppress microbiological activity. Under humid temperate conditions their part is probably subsidiary to the decompositions effected by micro-organisms.

HUMIFICATION.

We have seen under what conditions added organic matter becomes destructively oxidised. There is another class of processes whereby, although some destructive decomposition takes place, a considerable proportion of the added organic matter survives in the form of a dark-coloured amorphous material, which may be termed humified organic matter or humus. There are three distinct types of humification which will be examined in turn.

I. Anaerobic Humification. This type of decomposition is shown most typically in the changes undergone by the residues of aquatic vegetation in ponds and lakes. The gaseous products are methane, hydrogen, and carbon dioxide, and there is at the same time a production of amorphous dark material which, by continued accumulation,

forms fen or lacustrine peat. In ordinary soils, this type of decomposition is intermittent, being confined to those periods when soil aeration is defective, as during the winter months, or indeed at any time when the soil is very wet. It may be that, even when the general conditions in the soil favour aerobic decomposition, the conditions within the actual soil crumbs favour anaerobic decomposition.

The effect of temperature on anaerobic humification is somewhat masked by the fact that the higher temperatures of the summer months are generally associated with drier soil conditions, so that humification becomes recessive. But when, owing to subsoil water conditions or to heavy summer rainfall, the soil is kept in a permanently waterlogged condition, it may be presumed that the humification processes will be, equally with the oxidative processes, favoured by warmth. The development of peaty organic matter occurs, even in the tropics, in waterlogged situations.

Anaerobic humification goes on both in calcareous and non-calcareous soils under permanent waterlogging. When, however, waterlogging is only seasonal, the presence of calcium carbonate by promoting an open texture and free drainage, will tend to abridge the period of anaerobic decomposition.

The extreme result of anaerobic humification is seen in bog peat, the condition for whose formation is the prevalence of permanent wetness, either from the presence of a high water table, or from surface seepage, as in high moor and mountain peats. The origin and classification of peats will be examined in greater detail in a later chapter.

II. Acid Humification. This term may be applied to those decompositions affecting organic residues in the absence or deficiency of calcium and other bases. This is exemplified by the production of acid peat under heath conditions. That humification in this case is not conditioned by anaerobic processes may be shown by the fact that, on acid rocks such as quartzite, peaty organic matter is formed, even under conditions where aeration is most thorough.

Indeed, on such rocks it may be observed that the residues of crevice vegetation, such as lichens and mosses, readily form a peaty residue. This type of decomposition prevails under heath and natural coniferous forest. Conifers and heath plants make comparatively small demands on the mineral reserves of the soil and their residues are correspondingly poor in basic constituents. The decomposition of these residues is to a large extent carried out by fungi. The acid conditions preclude the activity of earth worms, and the peaty organic matter accumulates as a sharply defined layer at the surface.

In the literature of forestry, it is frequently stated that certain plants are associated with the production of peaty organic matter. The nature of this association is not clearly understood and is sometimes apprehended wrongly. Heather, *Erica cinerea*, and whortleberry, *Vaccinium myrtillus*, for example, are frequently mentioned as peat formers. Their association with the formation of peat may be attributed primarily to the circumstance that they make very small demands on the soil for mineral nutrients, and therefore tend to occupy soils which are naturally poor in such constituents, or which by degradation and impoverishment have fallen to a low mineral status. The low base status of their residues largely contributes to the development of peaty humus, the leachings from which still further deplete the scanty base reserves of the soils on which they are found.

Whilst it is impossible to decide whether the invasion of a peat-building vegetation has caused soil deterioration or whether these plants have occupied a soil which has become unfit to support the growth of plants with more exacting mineral requirements, it is certain that the production of an acid type of peat from the residues of peat forming plants intensifies the leaching of mineral matter from soils, and still further accentuates the conditions antecedent to its formation.

III. *Humification Under Semi-Arid Conditions.* The increasing aeration of the soil accompanying a decrease in

rainfall in passing from a humid to an arid climate is reflected in a progressive decrease in the proportion of organic matter in the soil under forest vegetation.

When the climate is no longer sufficiently humid to carry a natural forest vegetation, and the prairie or steppe region is entered, there is a marked increase in the organic matter content of the soil. This can be verified in any locality where, under the same rainfall, both steppe and forest vegetation occur. Moreover, in the steppe region, the highest contents of organic matter occur under more arid conditions than those of the zone of transition from steppe to forest.

It seems clear that the accumulation of organic matter in such soils, which include the black earths of Russia and the prairie soils of North America, must be the result of a type of humification which is essentially different from that which occurs in more humid climates. On the one hand, the scanty rainfall precludes the prevalence of anaerobic conditions, except in depressions, whilst on the other hand, the high base status prevents an acid type of humification.

Whilst the nature of the decompositions which occur under such conditions has not yet been completely investigated, it seems probable that the decisive factor is the occurrence of summer drought, resulting in a lowering of the soil moisture content below the point at which microbiological actions can proceed.

The high base status of the soils of the semi-arid regions also exerts an effect. It is noteworthy that, in the tropics, soils containing free calcium carbonate or a high content of exchangeable bases are usually dark in colour. It appears that a high lime status coupled with the occurrence of high temperatures results in a darker type of humus than that produced under more humid conditions. These conditions are reproduced in soils such as the prairie soils and the continental rendzinas which, although they do not occur in definitely arid climates, possess a high base status and are subjected to dry hot summers.

Owing to the fact that the type of humus formed under conditions of base-saturation is of a darker colour than the humus of base-unsaturated soils, a visual estimate of the organic matter content of soils of the rendzina and black earth groups may frequently lead to an exaggerated idea of the amount actually present. For example, in Barbados, red soils have a higher organic matter content than black soils derived from the same limestone material.

FACTORS GOVERNING THE ORGANIC MATTER CONTENT OF SOILS.

The amount of organic matter in a given soil represents the balance between the addition of material from plant residues and the destructive decomposition by soil micro-organisms. A low content of organic matter may result either from the scanty addition of plant residues, as in the case of desert soils, or from the rapid oxidative decomposition of organic material under the joint influence of high temperatures, free aeration, and a sufficiency of moisture for microbiological processes.

Broadly speaking, we may consider the humification processes as conservative of organic matter. The most rapid accumulation of organic matter in the soil will occur under perpetually waterlogged conditions, when the decomposition is mainly anaerobic.

High rainfall, by encouraging plant growth and maintaining soils in wet, and therefore inadequately aerated, conditions, favours the accumulation of organic matter. The relation between rainfall and organic matter content is, however, complicated by the effect of natural vegetation. Steppe or savanna is generally more favourable to the accumulation of organic matter than forest. Under forest conditions, the added residues consist principally of leaf-fall and this, as litter, is exposed to considerable losses by aerobic decomposition. Under steppe the residues are more thoroughly incorporated with the soil, and further microbiological decomposition is frequently inhibited through lack of moisture.

The inhibition of microbiological decomposition during the dry season may account for the considerable proportions of organic matter found in certain tropical soils. At present, little is known of the organic matter cycle under such conditions; but it is certain that cultivation and irrigation in hot climates lead to the rapid disappearance of organic matter from soils. Whilst it is not unusual to find as much as 10% of organic matter in virgin soils in the humid tropics, in irrigated soils of semi-arid regions, such as the Sudan, less than 1% may be present. Sugar cane soils in Trinidad under long cultivation may contain only 2-3% of organic matter in spite of heavy dressings of pen manure.

The relationships of organic matter to climate and cultivation in Great Britain are fairly simple and intelligible. Grassland soils are, as may be expected, richer in organic matter than arable soils, not only because they receive a greater contribution of organic matter from the residues of vegetation and the excreta of grazing animals, but also because the intense aeration produced by tillage operations favours oxidative decomposition by aerobic organisms.

Texture also has an influence on the organic matter content of soils. Light soils are generally better aerated and warmer, and the decomposition of added organic matter proceeds more rapidly than in close-textured soils.

Apart from texture, the water-air conditions are markedly influenced by situation and, in particular, by the position of the water table. With a high water table, even an open textured light soil may be maintained in a state of deficient aeration, with the resultant tendency for organic matter to accumulate as the products of anaerobic humification. Such conditions lead in extreme cases to peat formation.

The effect of calcium carbonate on the organic matter content under British conditions is somewhat complex. In the first place, the presence of free calcium carbonate confers a flocculated or crumb structure on heavy clay soils, and to that extent favours aeration and decomposition. Where

calcium carbonate becomes a dominant constituent, as in chalk soils, the texture is very open and the excessive aeration is further accentuated by the tendency to shallowness. It is not surprising, therefore, that calcareous soils and, in the highest degree, chalk soils tend to be deficient in organic matter. It is possible also that a high lime status in itself may be favourable to aerobic decompositions, apart from its indirect effect through soil texture.

Variations in climate have an important effect on the balance between gains and losses of organic matter. Under high rainfalls, as in the West of England and Wales, the soil is maintained in a wet condition through a long period of the year and, whilst vegetative growth is favoured, oxidative decomposition is considerably repressed. Further, the cool summers of the West result in markedly slower microbiological decomposition, even with equal aeration, than that which occurs in the hotter summers of S.E. England, with its approach to a continental climate.

NITROGEN CYCLE IN SOILS.

The organic matter of soils contains carbon and nitrogen in the approximate ratio of 10:1. Assuming that organic matter contains 55% of carbon and 5.5% of nitrogen, then, if the nitrogenous portion be assumed to be of a protein character, it would amount to $5.5 \times 6.25 = 34.4\%$, or a little over one-third of the total organic matter. There is, however, a distinction between the supposed protein of soil organic matter and fresh proteins, for whilst the latter, on being added to the soil, undergo rapid decomposition, the organic matter protein appears to be highly resistant to microbiological action, although it may be partially hydrolysed by dilute acids and decomposed by hydrogen peroxide.

There is in all soils a small proportion of soil nitrogen in the form of relatively simple compounds, such as amino-acids, amides, ammonium salts, and nitrates. It is from these compounds, and principally from ammonium salts and nitrates, that plants usually obtain their nitrogen.

It has long been known that nitrates are produced by the decomposition of nitrogenous organic matter. Indeed, until the discovery of the nitre deposits of Chile, the decomposition of organic matter was the main source of the supply of the potassium nitrate used for gunpowder manufacture. In 1877, T. Schloesing and A. Müntz⁴, under the impetus of the recent discoveries of Pasteur, showed that the production of nitrate from decaying organic matter was due to the action of bacteria. Later work, notably by R. Warington⁵, showed that the changes involved a successive simplification, according to the following scheme:—Nitrogenous organic matter→amino-acids and amides→ammonium salts→nitrites→nitrates. Whilst the changes from ammonium salts to nitrites, and from nitrites to nitrates were each shown to be the work of specific bacteria, *B. nitrosomonas* and *B. nitrobacter*, respectively, the process of ammonification was found to be the result of the work of many organisms, both fungal and bacterial, and possibly to include purely chemical changes. It should be added that fresh organic matter is much more susceptible to attack than humified organic matter, and is therefore the principal source of ammoniacal and nitrate nitrogen.

Nitrification, the production, via nitrites, of nitrates from ammonium salts, is an aerobic process, the work of autotrophic bacteria whose energy is obtained from the oxidation involved by this change. The minimum, optimum, and maximum temperatures for the process are about 5°C., 35-40°C., and 55°C. respectively. Whilst nitrification proceeds most rapidly in a neutral or slightly acid medium, it is now known that it can also go on, though more slowly, in soils which are definitely on the acid side of neutrality. The presence of moisture is, of course, essential, as for all microbiological actions.

The change, nitrites→nitrates is more rapid than the change, ammonium salts→nitrites; and for this reason nitrites in the free state do not occur in soils. The whole change from ammonium salts to nitrates is, under satisfactory

conditions for nitrification, more rapid than ammonification. The presence of ammoniacal nitrogen in amounts of more than a few parts per million of soil is evidence that conditions are unfavourable for nitrification. This is commonly the case in acid soils rich in organic matter. Such soils may contain up to 0.05% of ammoniacal nitrogen and be devoid of nitrates. Ordinarily, however, it may be said that the rate of nitrification depends upon the rate of ammonification.

The production of nitrates in soils is subject to marked seasonal influences. The low temperatures and the deficient soil aeration which obtain in the winter months result in an almost complete cessation of the process. The rate of production increases in the spring, and reaches its maximum in the summer months. The actual proportion of nitrate nitrogen, even in the same soil, varies enormously, for, not only is it subject to the demands of plant roots, but also, being in the form of soluble salts and not absorbed by the soil, it is rapidly washed out and lost in the drainage, if heavy rainfall occurs. There is also evidence that, apart from these circumstances, there are rapid fluctuations in nitrate content over short periods.

Whilst plants normally absorb their nitrogen in the form of nitrates, certain plants, among them some important crops such as potatoes, appear to prefer ammonium salts. In other cases, the nitrogen nutrition of plants is effected by symbiosis, as in the case of coniferous trees. The fibrous rootlets of such trees form a symbiotic association with the mycelium of certain fungi, to which the term *mycorrhiza* is applied. Mycorrhizal combinations are common on peaty acid soils and among plants of semi-xerophytic habit.

The nitrogen absorbed by the plant from the soil, either directly or through the intervention of mycorrhiza, is built up into complex plant proteins. These are returned to the soil again as residues, either directly or through the digestive tracts of grazing animals. The extent to which vegetable proteins consumed by human beings are returned as excreta depends on systems of sanitation. Under urban conditions,

the nitrogen of human food is almost completely lost to the soil.

The cycle, soil nitrogenous matter→ammonium salts→nitrites→nitrates→protein→soil nitrogenous matter is only one of the circulations involved in the complete nitrogen cycle. Apart from the nitrate absorbed by plants, there is a certain loss of nitrate from the soil in drainage. This may amount, under British conditions, to 10-60 lbs. per acre.

There is also the possibility that nitrates may be transformed by bacterial action into elementary nitrogen. Denitrification, as the process is called, represents a definite loss of nitrogen from the soil. It is the work of a bacterium, *B. denitrificans*, the conditions for whose activity appear to be the presence of large quantities of fresh organic matter and the alternation of aerobic and anaerobic conditions. It may be presumed that denitrification is of slight importance except under the highly artificial conditions of market garden soils.

NITROGEN FIXATION.

We have now to account for the stock of combined nitrogen present in soils. To refer it to the contribution of plant residues is only to change the problem to that of the source of the nitrogen supply of plants. At the outset, we may practically exclude the parent materials of soils as sources of soil nitrogen, since igneous rocks are devoid of this element, whilst sedimentary rocks only contain fractional proportions, which would be quite inadequate to account for the combined nitrogen of soils derived from them.

A possible source is the ammonia and nitric acid brought down by rainfall. This contribution amounts, under British conditions, to about 4 lbs. per acre per annum. Under tropical conditions, however, frequent and intense thunderstorms result in the formation of large quantities of nitric acid. It has been estimated that in Indo-China 30-35 kg. of nitric acid per hectare may fall annually in rainfall.

An annual accession of assimilable nitrogen of this magnitude, if shown to be general, might sufficiently account for the nitrogen nutrition of plants and the soil nitrogen of the tropics. Under the conditions obtaining in temperate and cold climates, however, it is necessary to examine other sources of supply.

In the early years of the nineteenth century, the great authority of Liebig was lent to the theory that plants obtained their nitrogen from the traces of ammonia present in the atmosphere, and that the use of nitrogenous fertilisers was as superfluous as the use of fertilisers to supply carbon. The careful experiments of Lawes and Gilbert at Rothamsted, however, clearly demonstrated the dependence of plants on supplies of available nitrogen in the soil. Leguminous plants were an apparent exception according to field experiments, which actually demonstrated a gain in soil nitrogen consequent on the growth of such crops. In practice, the favourable effect following the growth of leguminous crops had long been known, and is suggested by the familiar lines from Virgil's First Georgic.

In 1858, Lachmann⁶ showed that the nodules which had been observed in the roots of leguminous plants contained bacteria, which were finally isolated by M. W. Beijerinck⁷ in 1888, and are now known by the name *Bact. radiculicola*. These bacteria live in symbiosis with their hosts and have the ability to bring atmospheric nitrogen into organic combinations. The nitrogen thus fixed from the air is available for the nutrition of the host plant, from which in turn the bacteria draw their non-nitrogenous pabulum. They are aerobic organisms and appear to exist in forms associated with particular species of leguminous plants.

Certain other plants, notably the alder (*Alnus*), also exhibit nodule formation and form symbiotic associations with bacteria of the *Bact. radiculicola* group.

In addition to the nitrogen fixation effected by *Bact. radiculicola* in symbiosis with leguminous plants, there is another group of bacteria able to build up atmospheric

nitrogen into organic combination without the co-operation of a host plant. The free-living nitrogen fixers are heterotrophic organisms depending for their existence on a supply of soluble carbohydrates. The first species, discovered by S. Winogradsky⁸ in 1893, namely, *Clostridium pastorianum*, is an anaerobic organism originally isolated from pond mud. Under laboratory conditions, it can fix 2-3 mg. of nitrogen per gram of carbohydrate utilised. There is little doubt that bacteria of the *Clostridium* type play an important part in the nitrogen economy of soils with deficient aeration and even of normal soils during periods of deficient aeration.

In 1894, M. W. Beijerinck⁹ discovered an aerobic heterotrophic organism, *Azotobacter chroococcum*, which, like *Clostridium*, utilises soluble carbohydrates and fixes elementary nitrogen. It appears to be more efficient than *Clostridium* and fixes about 10 mg. of nitrogen per gram of carbohydrate oxidised. *Azotobacter* is more sensitive to acid reaction than *Clostridium* and ceases to function in media more acid than p.H. 6.0.

Other bacteria are known which can fix atmospheric nitrogen, and symbiotic associations between such organisms and algæ have been demonstrated. The direct fixation of nitrogen by higher plants, though occasionally reported, still lacks conclusive proof.

SULPHUR CYCLE IN SOILS.

Sulphur is present in soils mainly in the form of cystine groups associated with the proteins of organic matter. In addition, under anaerobic conditions, sulphides and even free sulphur may be present.

Certain bacteria in the soil have the ability to oxidise organic sulphur compounds, sulphides, and free sulphur to sulphuric acid or sulphates. These are aerobic bacteria whose activity appears to be favoured by the presence of calcium carbonate. Broadly speaking, it may be said that the same conditions which favour nitrification also favour the production of sulphates.

Under anaerobic conditions, sulphates are readily changed into sulphides, including hydrogen sulphide, whose odour is frequently evident in subaqueous horizons.

CARBON-NITROGEN RATIO OF SOIL ORGANIC MATTER.

A consideration of the composition of the plant materials added to the soil from the residues of vegetation will show that their ratio of carbon to nitrogen is in the region of 40 : 1. The organic matter of soils shows a much narrower carbon-nitrogen ratio. For the soils of temperate regions, the ratio approaches very closely to 10 : 1. Only a small proportion of the organic matter in soils consists of unaltered plant residues, and the carbon-nitrogen ratio observed represents the character of the actual soil humus. Sufficient data are not yet available for a satisfactory judgement to be formed as to the effect of climate on the C : N ratio. From a study of soils of the United States, H. Jenny¹⁰ concludes that the ratio widens with decreasing temperature, but W. McLean¹¹ has examined many tropical soils which give C : N ratios considerably greater than 10 : 1.

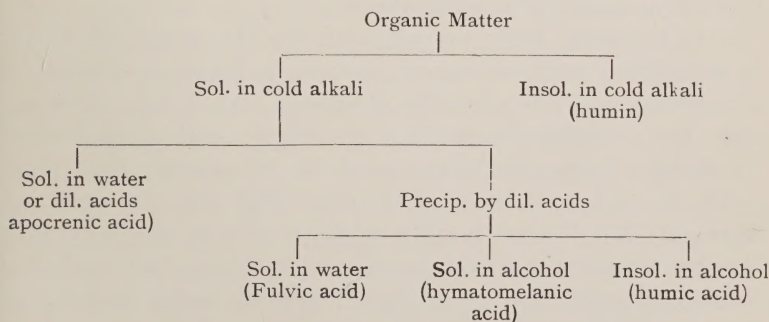
It might be expected that, since the greater part of the organic matter of the soil has resulted from the decomposition of plant remains, wide variations should occur, according to the character of the organisms which have operated and the composition of materials added from the vegetation or in the form of organic manures. Actually, however, this ratio varies within comparatively narrow limits. The addition to soil of organic matter with a wider carbon-nitrogen ratio than that characteristic for the soil is followed by a considerable evolution of carbon dioxide. The organisms concerned in this change require a supply of available nitrogenous compounds and there is thus a temporary diminution in the content of ammoniacal and nitric nitrogen. In the absence of fresh organic matter it may be supposed that losses of carbon and nitrogen from the soil by microbiological decompositions will be approximately in the proportion in which they occur in the characteristic organic matter of the soil.

THE CONSTITUTION OF SOIL ORGANIC MATTER.

The nature and constitution of soil organic matter has long presented a difficult problem for soil chemists. Its complexity was recognised at an early stage by the first workers in the subject. One of the chief difficulties in investigating the problem is the impossibility of isolating the organic matter of ordinary soils from its accompanying mineral matter. It was inevitable, therefore, that much of the earlier work was carried out on peats, consisting wholly or mainly of organic matter. And since the mode of origin of such organic matter is different from that in the majority of ordinary soils, it is probable that the results obtained are only of restricted application.

As far back as 1786, Achard extracted a black substance from peat by the action of alkali, and this gave the impetus to numerous attempts to isolate complex acids and other related compounds from peats and also from mineral soils. In this way, such compounds as humic acid, ulmic acid, crenic acid, apocrenic acid, humin, and ulmin were distinguished. But, inasmuch as they are non-crystalline substances of exceedingly complex composition, their identity as definite chemical compounds has always been open to doubt.

A possible fractionation of soil organic matter is shown in the following scheme:—



The very existence of humic and similar acids was called into question by certain workers, notably E. Gully¹², who

considered that the phenomena of extraction by alkalies and precipitation by acids were explainable as colloidal absorption effects.

The existence of acids in peat was placed beyond doubt by the work of S. Odén¹³, who showed that the conductometric curves obtained by adding increasing amounts of alkali to peat or prepared humic acids were such as could only be given by weak insoluble acids.

Odén concluded that humic acid has the formula $C_{60}H_{52}O_{24}(COOH)_4$. Its aqueous suspension has a pH of 4.5. It should be remarked, however, that pH values of less than 4 are frequently encountered in podsol profiles under heath peat.

O. Schreiner and E. C. Shorey¹⁴ succeeded in isolating a number of definite compounds from soil organic matter. Included in the humic acid fraction, they found paraffinic acid, agroceric acid, $C_{26}H_{44}O_3$, agrosterol $C_{26}H_{44}O$, and phytoserol, $C_{26}H_{44}O.H_2O$. In the fraction not precipitated by acids (crenic and apocrenic group) they found dihydroxystearic acid, α -picoline- γ -carboxylic acid, xanthine, hypoxanthine, cytosine, histidine, and arginine. No complete fractionation was effected and it is possible that some of the compounds isolated were not originally present as such in the soil, but were formed by the processes of extraction used by these workers.

It has long been known that, by the action of strong hydrochloric acid on sugar, various substances similar in appearance to humus can be obtained, and it was supposed that they might be similar in origin and constitution. Artificially prepared humic acid is, of course, free from nitrogen, whereas natural humic acid always contains nitrogen; but it was thought possible that it might be present in adventitious compounds which were not essential constituents of humus. It was, however, found possible to prepare artificial humic acids containing nitrogen by condensation of sugars with amino-acids. L. C. Maillard¹⁵ actually suggested this mode of origin in the soil.

V. H. Beckley¹⁶ demonstrated the formation of humus from carbohydrates by way of furfuraldehyde or methylhydroxyfurfuraldehyde, which undergo condensation to form humus.

W. Eller¹⁷ suggested an entirely different mode of origin by way of benzene derivatives. Humic acids prepared artificially by the oxidation of phenol derivatives were shown to resemble closely the humus of soil.

A more recent development is the theory that humic material in the soil originates from lignin, the complex molecule of which is built up from benzene ring derivatives. M. S. Du Toit and H. J. Page¹⁸, from a study of the humification of plant materials, were able to show a distinct correlation between loss of lignin and increase in humus. Whilst the lignin theory for the origin of organic matter may be adequate for many cases, it is improbable that lignin is the only source. Sphagnum, for example, the parent material of high moor peat, contains comparatively little lignin.

The dark colour of the humic matter of soils has tended to obscure the possibility that a considerable proportion of it may consist of materials which have survived from plant residues. R. A. Gortner¹⁹ found that alkaline extracts similar to those obtained from soils could also be obtained from unchanged vegetable materials. He suggested that the only characteristic soil constituent is a black pigment, probably of bacterial origin, which rarely exceeds 40% of the humus obtained by alkaline extraction. The pigment is poor in nitrogen, and soluble in aqueous ammonia.

S. A. Waksman²⁰, in the last of a series of papers which include a very full historical and critical account of the subject, suggests that the alkali-soluble humus of the soil consists of two main categories, namely, (a) lignin, which being unattacked by micro-organisms, persists from the added plant materials; and (b) microbial protoplasm and its decomposition products. The substances in the second category contain nitrogen.

S. A. Waksman and K. R. Stevens²¹ propose the application to soil organic matter of a system of proximate analysis used for plant materials. The results for the organic matter of some typical soils are shown in Table XIII.

TABLE XIII.—COMPOSITION OF ORGANIC MATTER FOR SOME TYPICAL SOILS. CALCULATED ON BASIS OF SOIL ORGANIC MATTER = C $\times$ 1.72 (WAKSMAN AND STEVENS).

SOIL	Ether Soluble	Alcohol Soluble	Hemi-Cellulose	Cellulose	Lignin Humus Complex	Protein
Summit soil, 4.49% org. matter ...	3.56	0.58	5.44	3.55	43.37	33.78
Kansas tshernosem, 5.98% org. matter	4.71	1.53	8.60	5.22	40.81	34.74
Alberta tshernosem, 11.20% org. matter	0.80	0.82	5.53	4.12	41.87	37.35
Carrington loam, 6.48% org. matter	0.62	0.62	8.21	3.64	42.29	30.38

Compared with plant residues such as straw and leaves, the soil organic matter contains much less of the hemi-cellulose and cellulose fractions. On the other hand, there is a considerable increase in the lignin fractions and the protein fraction. It appears highly improbable that the lignin shown in the above system of fractionation is identical with plant lignin, whilst the identity of organic matter protein with plant protein is even less probable. Waksman regards the soil lignin as principally surviving lignin, whilst the protein is mainly microbial in origin. Whatever be the nature of these fractions, they appear to account for 75% of the total organic matter of the soil.

The peats present a special type of distribution among the different fractions. Whilst in fen peats, cellulose is almost absent, in high moor (sphagnum) peats, considerable proportions of cellulose and hemi-celluloses appear to survive decomposition. Protein is considerably higher in low moor than in high moor peat.

FRACTIONATION OF ORGANIC MATTER BY HYDROGEN PEROXIDE.

G. W. Robinson and J. O. Jones²² suggested a simple fractionation of soil organic matter into humified and non-humified by the use of boiling 6% hydrogen peroxide as a reagent. The assumption underlying this method of determining the degree of humification has, however, been questioned by W. O. Robinson²³, who showed that, in the presence of soil, cellulose is appreciably attacked by 6% hydrogen peroxide. The peroxide method for determining the degree of humification of soil organic matter has, however, found a certain application in the study of forest soils. For example, A. Němec²⁴ was able to distinguish different types of forest humus by the use of this reagent.

W. McLean²⁵ has recently reopened the question of the attack of hydrogen peroxide on soil organic matter. Whilst it is undoubtedly true that 6% hydrogen peroxide, originally proposed for differentiating humified from non-humified or structural matter, has an appreciable attack on cellulose, a weaker peroxide (1-2%) is without action on this substance. By studying the effect of hydrogen peroxide of varying strength on the carbon and nitrogen content of soils, it was found that two stages of oxidation could be distinguished. Starting with the most dilute solutions, a complex of constant carbon-nitrogen ratio is progressively oxidised. Above a certain strength of peroxide, varying somewhat for different soils, there is no further oxidation of nitrogenous matter, but a progressive decomposition of non-nitrogenous compounds or complexes. The course of the oxidation of a Welsh brown earth is shown graphically in *Fig. 6*. It will be seen that, even with very dilute peroxide, there is a considerable oxidation of carbon and nitrogen; but whilst the oxidation of carbon slowly increases after about 86% has been attacked, there is no further attack of nitrogen beyond the first stage of attack.

The material oxidised in the first stage, termed by McLean the *oxidisable complex*, accounts for 70 to 80% of

the total carbon and nitrogen present, and is more readily oxidised in base-unsaturated than in base-saturated soils. It would appear to be identifiable with the lignin-humus complex and the protein of Waksman (*loc. cit.*). The non-nitrogenous organic matter oxidised in the second stage may represent cellulosic material which has not as yet undergone microbiological decomposition.

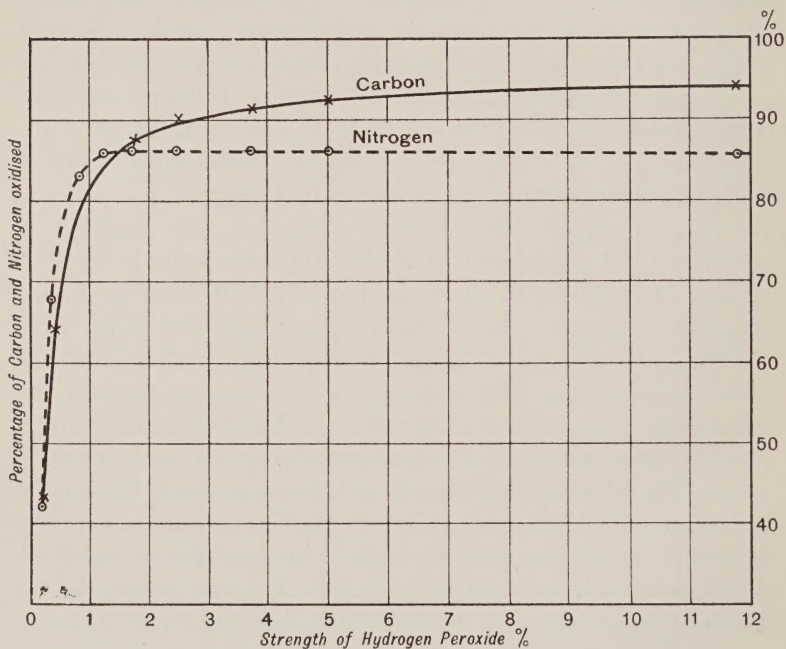


FIG. 6.—Oxidation of soil organic matter by hydrogen peroxide.

Whilst Waksman regards the lignin-humus and the protein as independent complexes, the one representing resistant material surviving microbiological decomposition and the other consisting mainly of altered microbial protoplasm, the constancy of the carbon-nitrogen ratio of the organic matter oxidised by weak peroxide does imply some close association between these groups. The high

resistance of humic nitrogenous compounds to attack by ammonifying organisms compared with that of known plant and animal proteins suggests that the nitrogen of the oxidisable complex may be built up in complexes such as the artificial "humus" prepared by L. C. Maillard (*loc. cit.*) from the condensation and carbohydrates and amino-acids.

The importance of lignin as a source of soil humus has been clearly demonstrated by the work of Waksman, Du Toit and others. But it cannot be assumed that Waksman's lignin fraction represents plant lignin. W. T. McGeorge²⁶ has shown that, whilst the lignin of plant materials possesses, in a certain degree, base exchange properties, the "lignin fraction" of soils has undergone changes resulting in a great increase in exchange capacity. To what extent these changes are biological is uncertain. Waksman considers lignin to be resistant to microbiological attack. But the rapid disappearance of plant residues by oxidative decomposition under certain conditions demonstrates their susceptibility to change. Indeed, M. Phillips, H. D. Weihe, and N. R. Smith²⁷ have shown that, under suitable conditions, lignin may be as rapidly decomposed by soil organisms as celluloses and proteins. And thus, even although the bulk of the non-nitrogenous portions of the humus may be derived from lignin, it is scarcely justifiable to assume that it persists as such. The work of McLean suggests rather that the humified organic matter of the soil, which may be tentatively identified with his "oxidisable complex," is a homogeneous constituent containing nitrogen as an essential ingredient.

BIBLIOGRAPHY.

- ¹ PRINGSHEIM, H., *Die Polysaccharide Berlin*, 1923; From S. WAKSMAN, *Principles of Soil Microbiology*, p. 428.
- ² DVÖRÁK, J., "Studien die Stickstoffanhäufung im Boden durch Mikro-organismen." *Z. landw. Vers. Stat. Oesterreich*, 1912, **15**, pp. 1077-1121; From S. A. WAKSMAN, *Principles of Soil Microbiology*, p. 428.
- ³ RUSSELL, E. J., "Oxidation in soils and its connection with fertility." *J. Agric. Sci.*, 1905, **1**, pp. 261-279.

- ⁴ SCHLOESING, T., and MÜNTZ, A., "Sur la nitrification par les ferments organisés." *Compt. rend.*, 1877, **84**, p. 301.
- ⁵ WARINGTON, R., "On nitrification." *J. Chem. Soc.*, 1878, **33**, pp. 44-51; *ibid.*, 1879, **35**, pp. 429-456; *ibid.*, 1884, **45**, pp. 637-672; 1891, **59**, pp. 484-529.
- ⁶ LACHMANN, "Über Knöllchen der Leguminosom." *Landw. Mitt. Zeitsch. K. Lehranstalt Vers. Stat.*, 1858; cited by S. WAKSMAN, *Principles of Soil Microbiology*, p. 124.
- ⁷ BEIJERINCK, M. W., "Die Bakterien der Papilionaceen-Knöllchen." *Botan. Zeitg.*, 1888, **46**, pp. 725-735, etc.
- ⁸ WINOGRADSKY, S., "Recherches sur l'assimilation de l'azote libre de l'atmosphère par les microbes." *Archiv. Sci. Biol. St. Petersburg*, 1895, **B**, pp. 297-352; from S. A. WAKSMAN, p. 107.
- ⁹ BEIJERINCK, M. W., "Über Oligo-nitrophile Mikroben." *Centrbl. Bakt.*, 1901, **II.**, **7**, pp. 561-582; from S. A. WAKSMAN, *Principles of Soil Microbiology*.
- ¹⁰ JENNY, H., "Relation of temperature to the amount of nitrogen in soils." *Soil Sci.*, 1928, **27**, pp. 169-188.
- ¹¹ MCLEAN, W., "The carbon-nitrogen ratio of soil organic matter." *J. Agric. Sci.*, 1930, **20**, pp. 348-354.
- ¹² GULLY, E., "Die Humussäuren im Lichte neuzeitlichen Forschungsergebnisse." *Int. Mitt. Bodenk.*, 1915, **5**, pp. 232-247 and 347-368.
- ¹³ ODÉN, S., "Die Humussäuren und die Bodenazidität." *Int. Mitt. Bodenkunde*, 1916, **6**, pp. 81-109.
- ¹⁴ SCHREINER, O., and SHOREY, E. C., "Chemical nature of soil organic matter." *U.S. Dept. of Agr. Bur. Soils Bull.*, 74, 1910.
- ¹⁵ MAILLARD, L. C., "Formation d'Humus et de combustibles minéraux par intervention de l'oxygène atmosphérique, des micro-organismes, des hautes températures ou des fortes pressions." *Compt. rend.*, 1912, **105**, pp. 1554-1556.
- ¹⁶ BECKLEY, V. H., "The formation of humus." *J. Agric. Sci.*, 1921, **11**, pp. 69-77.
- ¹⁷ ELLER, W., *Annalen*, 1923, **431**, pp. 133-186.
- ¹⁸ DU TOIT, M. S., and PAGE, H. J., "Studies on the carbon and nitrogen cycles in the soil." III.—The formation of natural humid matter. *J. Agric. Sci.*, 1930, **20**, pp. 478-488.
- ¹⁹ GORTNER, R. A., "The organic matter of the soil." I.—Some data on humus carbon and humus nitrogen. II.—A study of carbon and nitrogen in 17 successive extracts with some observations on the value of the black pigments of the soil. *Soil Sci.*, 1912, **2**, pp. 395-442, 539-548.
- ²⁰ WAKSMAN, S. A., "On the origin and nature of the soil organic matter or soil humus." V.—The rôle of micro-organisms in the formation of humus in the soil. *Soil Sci.*, 1926, **22**, pp. 421-436.
- ²¹ WAKSMAN, S. A., and STEVENS, K. R., "Contribution the chemical composition of peat." I.—Chemical nature of organic complexes in peat and methods of analysis. *Soil Sci.*, 1928, **26**, pp. 113-137.
- ²² ROBINSON, G. W., and JONES, J. O., "A method for determining the degree of humification of soil organic matter." *J. Agric. Sci.*, 1925, **15**, pp. 26-29.
- ²³ ROBINSON, W. O., "The determination of soil organic matter." *J. Agric. Res.*, 1927, **34**, pp. 339-356.

- ²⁴ NĚMÉC, A., *Proc. Int. Soc. Soil Sci.*, **2**, pp. 255-257.
- ²⁵ McLEAN, W., "Effect of hydrogen peroxide on soil organic matter." *J. Agric. Sci.*, 1931, **21**, pp. 251-261; "The nature of soil organic matter as shown by the attack of hydrogen peroxide," pp. 595-611.
- ²⁶ McGEORGE, W. T., "The base exchange properties of soil organic matter." *Arizona Agric. Techn., Bull.* 30, 1930, pp. 181-213. "Organic compounds associated with base exchange." *Arizona Agric. Techn. Bull.* 31, 1931, pp. 215-237.
- ²⁷ PHILLIPS, M., WEIHE, H. D., and SMITH, N. R., "The Decomposition of lignified materials by soil micro-organisms." *Soil Sci.*, 1930, **30**, p. 383.

Note.—S. A. Waksman and K. R. N. Iyer (J. Washington Acad. Sci., 1932, **22**, 41-50) have recently made an important contribution to the knowledge of soil organic matter by synthesising a product from lignin and protein which closely resembles the alkali-soluble humus of the soil in its biological and chemical reactions. For this substance, which is described as a ligno-proteinate, the term *humus-nucleus* is proposed. There is strong evidence for believing the humified organic matter of soils to be of this character.

CHAPTER VII.

GENERAL PHYSICAL PROPERTIES OF SOILS.

SOILS exhibit certain peculiarities in their physical properties by reason of the fact that they are essentially mixtures of mineral particles of varying size, together with closely associated colloidal clay and colloidal organic matter. The physical character and behaviour of a soil are governed by a number of circumstances, namely :—

(1) The nature and the size distribution of the particles composing the non-colloidal skeleton or framework.

(2) The content of colloidal material.

(3) The character of the colloidal material, depending on (a) the relative proportions of organic and inorganic colloidal material; (b) the constitution of the inorganic colloidal material; and (c) the nature and content of the exchangeable bases present.

(4) The degree of dispersion of the soil, considered as a colloidal system, in other words, the extent to which the soil is aggregated into crumbs.

(5) The moisture content.

TRUE AND APPARENT DENSITY.

It is necessary to distinguish, in the first place, between the true and the apparent density of a soil. Whilst the true density is an additive function of the densities of the soil constituents, the apparent density will vary according to the state of aggregation of the soil and the proportion of the apparent volume occupied by the interstitial spaces which are present even in the most compact soil.

The true density of a soil depends principally on the relative proportions of inorganic and organic matter present;

for, apart from certain heavy minerals, such as magnetite, which rarely occur in appreciable quantities, the densities of the inorganic constituents of the soil fall within a fairly narrow range. The following are the densities of the principal mineral ingredients of soils:—

Quartz	2.6 -2.7	Amphiboles & Pyro-	
Orthoclase	2.54-2.57	xenes	2.90-3.6
Albite	2.605	Olivine	3.2 -3.5
Oligoclase	2.65	Muscovite	2.76-3.1
Labradorite	2.68-2.71	Biotite	2.8 -3.2
Anorthite	2.765	Chlorite	2.6 -2.96
Apatite	3.16-3.22	Hæmatite	4.9 -5.3
Magnetite	4.9 -5.2	Limonite	3.6 -4.0
		Clay	2.6 -2.7

The specific gravity of soil organic matter varies from 1.2 to 1.7, but the figure is open to some doubt since the organic matter of ordinary soils cannot be isolated unchanged.

Since by far the greater part of the mineral matter of ordinary soils consists of quartz, feldspars, and clay, the specific gravity of the mineral soil free of organic matter is generally about 2.65. It is lowered by the presence of organic matter, and raised by the presence of oxides of iron such as magnetite, hæmatite, and limonite, and, to a smaller degree, by the presence of ferromagnesian minerals and micas. It is generally assumed that the specific gravity of a soil is an additive function of the specific gravities of its constituents.

The true density of a soil has proved of comparatively little value for diagnostic purposes, since the presence of notable proportions of organic matter or hydrated ferric oxide is readily detected by other means. It must be admitted, however, that no attempts have been made to correlate density with soil constitution over the very wide range of soil which it is necessary to consider in modern investigations. If such correlations were worked out, the density might prove to have a certain significance and serve as an aid to diagnosis where other methods of investigation are impracticable.

PORE SPACE.*

If a certain volume of a soil in its natural conditions be considered, it is evident that only a proportion of that volume is occupied by the soil particles. The remainder consists of interstitial spaces which, under ordinary field conditions, are occupied partly by water and partly by air. The weight of unit volume of dry soil, including the interstitial spaces, gives the apparent density.

The relationship of the true density, d , and the apparent density, d_1 , to the pore space will be apparent from the following considerations. If the percentage of pore or interstitial space be P , then the true volume of 100 c.c. of soil will be $100-P$, and the weight $(100-P)d$. The apparent density will therefore be $(100-P)d \div 100$, or $d-Pd/100$. The pore space expressed in terms of true and apparent density is given by $P = 100(d-d_1)/d$.

F. H. King¹ gives the following figures for the percentage pore space of typical soils:—

Sandy soil	32.5	Loamy clay	45.3
Loam	34.5	Clay loam	47.1
Heavy loam	44.1	Heavy clay	52.9

Such figures should only be regarded as approximations, for wide variations may occur in the same soil with changes in aggregation. Pore space is increased by addition of organic matter, and may rise in grass soils to 60%, or more.

Taking 50% as the pore space under natural conditions, it is apparent that this does not represent the closest possible packing of the particles of which the soil is composed. With uniform spheres, the tightest packing gives a pore space of about 26%. Where particles of different size are present and can pack into the interstices between the larger particles, the pore space can be considerably reduced. It is evident that, with the particle range of ordinary soils, the

*The subject matter of this section has already appeared in Chapter II, but is given again here for the sake of completeness.

tightest possible packing would result in a pore space very much less than 50%. The high figure actually obtained is due to the aggregation of particles into compound particles or crumbs. Any treatment which destroys the crumb structure and results in the soil particles behaving singly leads to a decrease in pore space. This change may be brought about mechanically by trampling, or by the beating of heavy rain, and chemically by the action of certain substances known as deflocculants, of which sodium carbonate is the chief example. The change of a saline soil into an alkaline soil consequent on the removal of soluble sodium salts is accompanied by deflocculation, involving decrease in pore space.

Pore space generally decreases from the surface downwards, but may pass through a minimum in an illuvial horizon, particularly if there is compaction. Unfortunately, its determination in profiles is beset with many difficulties.

TOTAL SURFACE OF SOILS.

The subdivision of the soil material implied by its granular character involves a great development of surface, which is the more marked the finer is the texture of the soil. This is clear from the following considerations.

A cube having sides 1 cm. in length has a total surface of 6 cm². If, now, the cube be subdivided into millimetre cubes, the total surface will become $1000 \times 6(0.1^2)$ cm². or 60 cm². The general relationship among particles of similar shape is that the specific surface, i.e., the surface per unit weight, is inversely proportional to the linear dimensions, i.e., in the case of spheres, inversely proportional to the radius.

It is possible, from a consideration of the mechanical composition curves of soils, to obtain an estimate of their specific surface.

The total surface of unit weight of a uniform soil fraction of diameter D , assuming the particles to be spherical, is equal to:—

Number of particles $\times$ surface of each particle

$$= \frac{1}{vd} \times \pi D^2$$

[Where v = volume of each particle, d = density, for which the value 2.65 may be assumed, D = diameter of particles.]

$$= \frac{1}{\frac{\pi D^3}{6} \times 2.65} \times \pi D^2$$

$$= \frac{2.264}{D}$$

If 100 g. of soil consist of $p_1, p_2, p_3 \dots p_n$, per cent. of particles having diameters $D_1, D_2, D_3 \dots D_n$, respectively, the total surface is equal to

$$2.264 (p_1/D_1 + p_2/D_2 + p_3/D_3 + \dots p_n/D_n)$$

A measure of the total surface can be obtained graphically by the method of G. Krauss² depending on the following argument.

Let the composition of the soil be set out in the form of a number of sectors, in which the angle of each sector is proportional to the percentage, p , of each fraction, and the radius proportional to $1/\sqrt[4]{v}$, the reciprocal of the fourth root of the settling velocity.

Since the area of a sector, S , is proportional to the angle, and to the square of the radius,

$$S = k_1 p / \sqrt{v} \quad (k_1 = \text{a constant})$$

But, from Stokes' Law, $D \propto \sqrt{v}$,

$$\therefore S = k_2 p / D \quad (k_2 = \text{a constant})$$

The area of unit angle of each sector is thus inversely proportional to the diameter and, consequently, proportional to the total surface of the fraction represented by it; and the sum of the areas of the sectors is proportional to the total surface of the soil.

In view of the smooth character of the mechanical composition curves of soils, it is sufficient to mark off at angular

intervals, equivalent to 10%, radii proportional to the reciprocal of the fourth root of the corresponding settling velocity, beginning with the lowest settling velocity. The

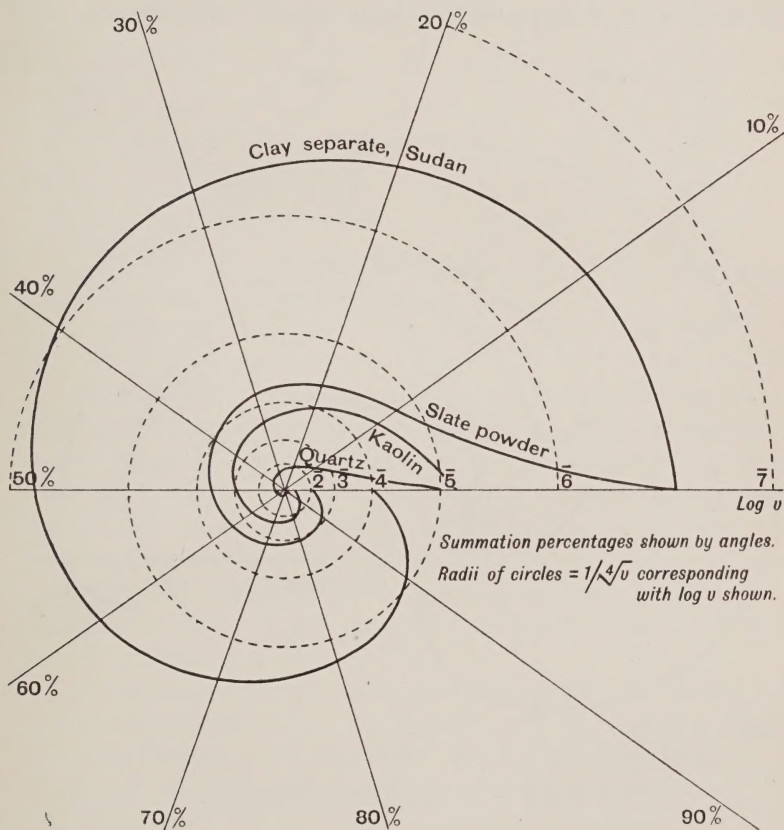


FIG. 7.—Total surface of certain materials.

area obtained by joining up the extremities of the radii is evaluated by means of a planimeter.

Curves drawn in this manner are shown in *Figs. 7 and 8*. The data are taken from the mechanical composition curves shown in *Figs. 1 and 2*.

The areas in square centimetres on the scale used and the estimated surfaces are shown in Table XIV. The latter figures are calculated on the assumption that the total surface per gram of uniform spherical particles, 0.002 mm. in diameter, is $2.264/0.0002 = 11320$ square centimetres, and

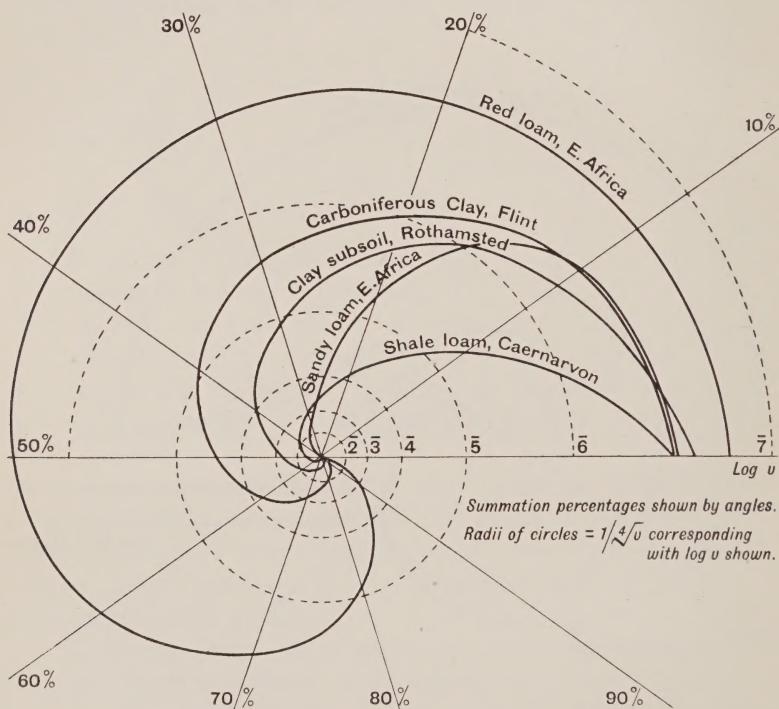


FIG. 8.—Total surface of certain soils.

that such particles have a settling velocity equivalent to 10 cm. in 8 hours.

Inspection of the curves will show at once the small total surface possessed by mechanically comminuted materials such as slate powder and quartz powder, even although they contain very little coarse material. The small contribution of the coarser fractions to the total surface is also evident.

TABLE XIV.—TOTAL SURFACE OF TYPICAL MATERIALS.

	Area included by curve sq. cm.	Estimated surface, square cm. per gram.
Quartz powder	1.16	1913
Kaolin	7.80	12860
Slate powder	15.87	26130
Shaly light loam, Caernarvon...	18.45	30400
Sandy red loam, East Africa ...	36.97	60890
Clay subsoil, Rothamsted ...	43.48	71690
Raw Carboniferous clay, Flint	60.31	99350
Clay separate, Sudan ...	118.5	195200
Heavy red loam, East Africa...	164.4	271800
Uniform fraction, 0.002 mm. diameter (calculated) ...	6.86	11320

The estimates thus obtained are probably no measure of the actual specific surface of the soils to which they refer, since under natural conditions a large proportion of the colloidal material, mainly responsible for these large figures, is probably built up into loose structures which become broken down in the operations of analysis. The values obtained may, however, serve a useful purpose in giving a comparative view of the degree of dispersion of which the specific surface is a measure.

From measurements of hygroscopicity, E. A. Mitscherlich³ obtained values for the specific surface, varying from 138 cm². for fine sand to 96.670 cm². for a heavy Java clay. These are now generally agreed to be over-estimates.

HEAT RELATIONSHIPS OF SOILS.

The temperature of the soil is closely related to the temperature of the air. But, whilst the mean temperature of the soil throughout the year approximates very closely to the mean temperature of the air, the seasonal and diurnal variations show different courses according to the depth below the surface. Below a certain depth, the temperature

of the subsoil is no longer affected by seasonal and diurnal changes and shows the ordinary increase with depth due to the influence of the earth's internal heat.

The response of the soil to additions or losses of heat in the form of radiation depends mainly on its specific heat. Whilst the specific heat of dry soil shows a certain amount of variation, far greater variations are due to changes in moisture content, owing to the high specific heat of water. The effect of varying moisture contents on the specific heat of soils is illustrated by the data from E. A. Mitscherlich⁴ in Table XV.

TABLE XV.—VOLUME SPECIFIC HEAT (CALORIES PER C.C.) OF SOILS WITH VARYING MOISTURE CONTENTS (MITSCHERLICH).

Soil	Volume specific heat in calories per c.c. at varying degrees of saturation		
	Dry	50% Saturated	Saturated
Sand ...	0.302	0.510	0.717
Humus ...	0.148	0.525	0.902
Clay ...	0.240	0.532	0.823

In considering the specific heat of soils, it is generally desirable to express it on a volume rather than on a weight basis. From the point of view of field conditions, this leads to a truer estimate of the response of the soil to gains or losses by radiation than that obtained by the ordinary method of expression.

Soils which are habitually moist respond more slowly to the heating effect of solar radiation than dry well drained soils. And thus, in the spring, the beginning of active plant growth takes place earliest in the driest soils.

The character of the diurnal variations in soil temperature is seen from the following data obtained by Mutterich⁵ in a sandy soil in Eberswald, Germany, in June, 1889. (Table XVI.)

TABLE XVI.—AIR TEMPERATURE, AND SOIL TEMPERATURE AT VARYING DEPTHS (MUTTERICH).

Time	Temperature °C.				
	Air	Surface	15 cm. depth	30 cm. depth	60 cm. depth
12 midnight	13.80	16.71	19.42	17.90	15.88
2 a.m. ...	12.90	15.59	18.42	17.59	15.91
4 a.m. ...	12.53	15.14	17.84	17.33	15.91
6 a.m. ...	14.68	15.89	17.35	17.03	15.92
8 a.m. ...	17.99	17.54	17.52	16.75	15.01
10 a.m. ...	21.05	22.65	18.72	16.59	15.89
12 noon ...	21.97	25.00	20.52	16.64	15.84
2 p.m. ...	22.61	26.37	22.08	17.00	15.80
4 p.m. ...	22.38	25.89	22.91	17.37	15.77
6 p.m. ...	21.24	22.32	22.64	17.71	15.75
8 p.m. ...	17.55	19.76	21.65	17.99	15.77
10 p.m. ...	14.80	17.81	20.53	18.06	15.88
Mean ...	17.79	20.06	19.97	17.33	15.85

The actual temperatures observed under other conditions will vary according to the course of the air temperature, the intensity of the solar radiation during the day, and the rate of loss by radiation during the night: yet the data in the above table illustrate the main features of the temperature changes in the soil. In sunny weather, the temperature of the actual surface of bare soil reaches a higher maximum than the air. Indeed, surface temperatures as high as 85°C have been reported from the tropics. Below the surface, the Eberswald data show that, with increase of depth, the amplitude of the diurnal temperature changes decreases, whilst the maxima and minima of the curve show an increasing lag behind the maxima and minima on the air temperature curve. Thus, whilst in the air, the maximum is at 2 p.m., and the minimum at 4 a.m.; at 15 cm. the maximum is at 4 p.m. and the minimum at 6 a.m.; and at 60 cm. the maximum is at 6 a.m. and the minimum at 6 p.m.

The corresponding amplitudes are 7.81° , 5.56° , and 0.17° , respectively.

Under a cover of grass vegetation, the amplitude of the temperature changes at the surface is less than in bare soil, whilst still smaller variations occur under forest. But in every case there is the same tendency for amplitude to decrease with depth and for the changes to lag behind those at the surface.

Situation and aspect affect the temperature of the soil. For each 300 feet increase in altitude there is a fall of approximately 1°F in air temperature, which is reflected also in the soil temperature. South slopes, in the Northern Hemisphere, are warmer than north slopes since they receive more intense insolation.

Colour affects soil temperature. Dark coloured and red soils absorb solar radiation better than light coloured soils. Wollny, comparing the temperature, at 4 inch depths, of soils covered with black and white material, respectively, found differences of nearly 6°C in favour of the former.

The seasonal changes in temperature show a similar tendency. With increase in depth, the difference in mean temperature between summer and winter becomes smaller and finally disappears. From A. A. Rambaut's⁶ observations at Oxford, it appears that, under a grass vegetation, the temperature at 10 feet varies from a minimum of about 45.5°F in March to a maximum of 56°F in September; at 6 feet the range is from rather less than 44°F in February to 59°F in August; whilst at 6 inches the range is from 38.5°F in February to 66.5°F in July. It is thus evident that the temperature of the soil lags markedly behind that of the air, and that lag increases and amplitude decreases with the depth.

Whilst variations in soil temperature are due mainly to the seasonal and diurnal changes in solar radiation, minor fluctuations are caused by other circumstances. Clouds check the loss of heat from the soil by radiation during night and hinder access of heat by radiation during day. Drying winds blowing over moist soils cause a lowering of tempera-

ture corresponding with the latent heat required for the evaporation of moisture. Condensation in cold soil from moist warm air results in gains of heat.

SHRINKAGE AND EXPANSION.

Anyone familiar with a clay country will have noticed the cracking of the surface of the soil during drought. The shrinkage on drying which leads to this cracking is most marked in the heaviest clays. The fissures thus opened often extend several feet into the soil and result in a further intensification of the drying out process. Material from the surface sometimes falls into the fissures and it has been shown by H. Greene⁷ that, in the Sudan Gezira, this leads to a considerable circulation of soil material.

Shrinkage on drying is also observed in highly organic soils. The drying out consequent on the drainage of peat soils is accompanied by marked contraction, so that the surface may be appreciably lowered.

Under laboratory conditions shrinkage is studied by moulding soil with water into a plastic paste and measuring the volume occupied at varying moisture contents during drying. In such a paste, the volume is represented by the soil and water only, for, by the method of preparation, interstitial air spaces, such as occur under natural conditions, are excluded. The curve connecting volume with moisture content is at first linear, the decrease in volume corresponding exactly with the loss in moisture. As the paste becomes drier, however, the curve begins to depart from the initial straight line and the diminution in volume lags behind the loss in moisture. At this point interstitial air spaces make their appearance and there is a change in colour towards that of the dry soil. The shrinkage which occurs after the linear relationship has been departed from is termed *residual shrinkage*.

W. B. Haines⁸ has determined the residual shrinkage of materials formed into plastic blocks. His results are shown in Table XVII.

TABLE XVII. — SHRINKAGE OF PLASTIC BLOCKS AND CLAYS
(HAINES).

Soil		Total Shrinkage %	Residual Shrinkage %	Clay % (log $v=4$)	Ignition Loss %
Clay separate	...	130.0	8.6	90.5	9.5
Kaolin	...	30.0	nil	52.8	12.4
Sudan clay	...	67.5	7.6	42.3	11.4
Clay subsoil	...	34.0	2.0	33.8	9.6
Harpenden common		37.5	10.3	25.5	10.5
Rothamsted loam	...	21.0	2.2	14.8	8.3
Sandy soil...	...	9.5	2.6	5.3	8.3
Peat soil	...	85.5	18.2	—	27.0

Whilst the total shrinkage represents roughly the water holding capacity of each material, the residual shrinkage appears to be correlated with the proportion of colloidal gel material present. It varies from nil in the case of kaolin to 18.2% in the case of a peat soil. In the case of kaolin or any other gel-free powder, shrinkage corresponds with water loss until the grains come into contact, after which no further shrinkage can occur. But where gel material is present, it is interposed between the mineral grains and opposes a resistance to shrinkage in the later stages of drying. The property of residual shrinkage can be conferred on non-colloidal powders such as kaolin by admixture of colloidal material such as gelatin.

The results obtained in the laboratory are only partially representative of field conditions, where, even in wet soils, a certain proportion of air spaces is always present above ground-water level. The laboratory conditions are most closely simulated by the individual "clods," which, in varying proportions, are always present in clay soils.

The shrinkage of soils on drying produces certain characteristic structures such as prismatic, columnar, and nuciform, which are often of considerable diagnostic value in the study of profiles.

PLASTICITY.

Plasticity may be defined as the ability of a material to change shape continuously under the influence of an applied stress, and to retain the impressed shape on the removal of the stress. The exhibition of this property by soils is dependent on the presence of certain proportions of water. A. Atterberg⁹ has distinguished an *upper* and a *lower plastic limit*. The upper limit of plasticity of a soil is the point at which the soil-water paste is just fluid, i.e., the point at which it fails to retain the shape impressed on it by an applied force. The lower limit of plasticity is the point at which cylindrical portions of a soil-water paste crumble down on being rolled, i.e., the point at which the paste is unable to change shape continuously under the influence of an applied force. The difference between the moisture contents, calculated on the dry soil, at the upper and lower limits is termed the *plasticity number*.

In Table XVIII, the plasticity data obtained by Atterberg for some typical Swedish soils are given.

TABLE XVIII. — PLASTICITY DATA FOR SOME TYPICAL SOILS (ATTERBERG).

Soil		Upper Plastic Limit % Water	Lower Plastic Limit % Water	Plasticity Number
<i>Highly Plastic Soils—</i>				
Silurian clay	...	67	40	27
Ancylus clay	...	57	30	27
Glacial clay	...	51	26	25
<i>Moderately Plastic Soils—</i>				
Fresh water clay	...	52	37	15
Arable soil	...	42	30	12
Glacial clay	...	32	25	7
<i>Feebly Plastic Soils—</i>				
Arable soil	...	64	58	6
" "	...	33	28	5
" "	...	22	18	4

Plasticity is associated with the presence, in soils and clays, of inorganic colloidal material. The property can be exhibited also by finely ground lamellar minerals such as mica and biotite. It is more strongly developed in soils with siliceous clay fractions than in soils whose clay fractions contain notable proportions of sesquioxides.

COHESION.

Cohesion is closely associated with plasticity, for plastic soils are cohesive and set into hard clods on drying. It is markedly affected by changes in moisture content and, in heavy and medium textured soils, increases as the soil dries. The intractable clods formed in the drying out of heavy clay soils are familiar to cultivators. With extremely light sands, however, cohesion may pass through a maximum with decreasing water content and then decrease as complete dryness is approached. This is exemplified by the looseness of sandy soils when completely dry.

Two types of cohesion may be distinguished. At low moisture contents, the effect is due in the main to forces which are considered by F. Hardy¹⁰ to be of a chemical nature, depending on the presence of active atoms or atomic groups having marked residual affinity. At higher moisture contents, cohesion is mainly due to film tension effects.

S. Johansson¹¹, using a method devised by Atterberg for the measurement of cohesion, has constructed curves connecting cohesion with moisture content for typical soils. The curves appear to be composite in character, but the work of W. B. Haines¹² throws some doubt on the discontinuous character of the cohesion-moisture relationship.

COLOUR.

The colour of the soil, especially the changes in colour through the horizons of the soil profile, is one of the most important aids to the recognition and description of the different genetic groups.

In the first place, it should be remarked that there is always a progressive lightening in colour as a soil dries out. The change may be attributed to the development of air spaces in the soil crumbs or in the colloidal coating of the coarse particles and, finally, to the partial dehydration of the colloidal material itself. Until a soil has dried down to the point at which air spaces occur there is, thus, no change in colour. In laboratory experiments, the beginning of colour change is found to correspond with the point at which there is a change in the curve connecting volume with moisture content. When the soil has reached the stage of desiccation at which only hygroscopic moisture is present there is no further change in colour.

Broadly speaking, it may be said that the most marked differences between wet and dry soil colours are found in soils containing high proportions of colloidal material. For example, black peaty soils may become light grey when dry.

The colours of moist soils are generally more vivid and strongly contrasted than those of dry soils. Soils which stand out sharply from each other under moist field conditions are much less distinct when they become dry laboratory samples. It is therefore preferable to use the colour of the moist soil for descriptive purposes where possible. The drying to which soils are subjected in the preparation of laboratory samples appears to involve colour changes which are not readily reversible on re-moistening.

The colour of the moist soil is governed principally by two constituents, namely, the colloidal organic matter and the colloidal clay. Generally speaking, organic matter confers the dark colour on soils. Among soils of the same group, the colour becomes darker with increasing proportions of this constituent. It should be noted that the effect of organic matter is more marked in sandy than in silty or clay soils. A coarse sand with 10% of organic matter may be nearly black when moist, whilst the same proportion of organic matter in a clay soil would have only a slight effect on the mineral colour.

The darkening effect of organic matter varies among the great soil groups. In the tshernosems, the rendzinas, and certain tropical and sub-tropical soils with high lime status, the humification of plant residues appears to produce a very dark coloured humus. On the other hand, in the red soils of the humid tropics, the humus appears to be lighter in colour. It may be, however, that its colour is masked by the highly coloured clay colloids of such soils. In the bleached horizon of podsol soils the organic matter would appear to be only slightly coloured. That differences exist in the colour of humus is only to be expected from the diversity of the processes whereby it is produced.

Except in highly organic soils, the colour is governed mainly by the soil mineral matter and, in particular, by the colloidal clay. In studying soil colour, it is generally convenient to destroy the humus by boiling with sufficient 6% hydrogen peroxide, whereby the true mineral colour may be observed. The removal of even small proportions of organic matter by this treatment results in an appreciable "brightening" of colour, whilst in the case of dark coloured organic soils the colour change is frequently remarkable.

Considering the mineral colour of soils, we may say at the outset that the unweathered minerals of the soil contribute but slightly to soil colour, which is conferred mainly by the inorganic weathering complex or clay complex. H. E. Annett¹³ has shown that finely divided magnetite is largely responsible for the dark colour of certain black cotton soils in India. Blue and green colours in subsoils may be attributed to the presence of ferrous compounds such as pyrites, marcasite, vivianite, and glauconite.

The mineral colour is mainly due to compounds of iron, and to a less extent, of manganese. Yellow, brown, and red colours are due to the presence of ferric oxides with varying degrees of hydration. Broadly speaking, red colours are associated with lower degree of hydration, such as are commonly encountered in tropical soils. Some effect on colour must, however, be attributed to the stage of aggregation of

the hydrate of ferric oxide present. The effect of a coloured constituent is more evident in sandy soils than in soils of heavier texture. On account of the low specific surfaces of such soils, comparatively small proportions of hydrated ferric oxide are sufficient to confer a marked red or brown colour. The strong colours of certain soils of the Trias and Old Red Sandstone formations and of certain desert soils do not therefore necessarily denote high proportions of ferric oxide. The most vivid red colours are found in soils of the humid tropics. Red colours may also occur in desert soils. High temperatures are apparently necessary for their development, which may be due to decomposition of the clay complex.

Soil colour is markedly influenced by the constitution of the clay complex. Generally speaking, soils in which the weathering complex has originated in the presence of notable proportions of calcium compounds, such as the tshernosems, rendzinas, chestnut earths, and grey desert soils, show, so far as the mineral portion of the soil is concerned, grey or greyish brown colours. The clay complex of such soils is rich in silica, and it may be presumed that sesquioxides exist only in combination with silica and not in the free state.

Grey colours occur in the A horizons of podsol profiles, due to the removal of ferric oxide by acid percolating waters. The grey colour is often only evident after removal of humus by peroxide treatment.

Grey colours are found also in the A horizons of soloti or degraded alkali soils. In such soils, the leaching out of the sodium soil in the absence of calcium carbonate leads to a decomposition of the clay complex and an eluviation of sesquioxides, leaving an A horizon rich in residual silicic acid.

There is also a large group of soils in which, although notable proportions of iron are present, grey colours prevail owing to reduction to ferrous compounds. These are the soils of marshes and bottoms where free aeration is prevented

by excess of moisture. Such soils occur not only in cool and temperate regions, but also in the tropics, where the grey mineral colour is frequently masked by dark organic matter.

In considering the colour of soils it is important to decide whether the colour reflects recent pedogenic process or whether it is simply a rock colour. The red colour of the soils of the English Trias and Old Red Sandstone must be attributed to the parent rock and not to the operation of contemporary processes. It represents rather the result of the weathering processes of Triassic and Devonian times.

SPECIFICATION OF SOIL COLOURS.

The accurate specification of the colour of soil is a matter of considerable difficulty both on account of the great subjective errors to which an estimate of colour is liable and also on account of the lack of any generally recognised scale of colours.

Numerous attempts have been made to arrive at a system of recording soil colours. The soil colours may be matched with a colour of known specification in a colour atlas; or, alternatively, they may be matched with composite colours, obtained by the combination, in varying proportions, of standard colours. The method employed by J. G. Hutton¹⁴ may serve as an example.

The desired soil colour was obtained by combination, in a rapidly rotating disc, of varying proportions of four standard colours, namely,

Neutral 9	...	...	...	...	...	White
Neutral 1	...	...	...	...	...	Black
Red 4/9	...	...	...	...	...	Red
Yellow 8/8	...	...	...	...	...	Yellow

The colours were obtained from the Munsell Colour Co., of Baltimore, Md., and were accurately specified by the U.S. Bureau of Standards.

A selection from Hutton's data for the colour value of some typical soils is shown in Table XIX.

TABLE XIX.—COLOUR VALUE OF TYPICAL SOILS (HUTTON).

Soil Colour Name	White %	Black %	Yellow %	Red %
Grey	22.0	60.0	11.0	7.0
Dark grey	16.0	67.0	9.0	8.0
Yellow	11.5	19.0	45.5	24.0
Greyish brown	14.0	50.0	22.0	14.0
Reddish brown	7.0	44.0	18.5	30.5
Dark reddish brown	3.0	72.5	10.5	14.0
Light loam	10.0	59.0	18.0	13.0
Dark brown	4.0	73.0	12.0	11.0
Dark greyish brown	9.0	73.0	11.0	7.0
Red	3.0	35.0	20.0	42.0
Dark red	0.0	70.0	7.0	23.0

BIBLIOGRAPHY.

- ¹ KING, F. H., *Physics of Agriculture*, Madison, Wis., 1910.
- ² KRAUSS, G., "Neue methode der mechanischen Bodenanalyse, sowie ein einfaches graphisches Verfahren zur Bestimmung der Körnoberfläche." *Int. Mitt. Bodenk.*, 1923, **13**, pp. 147-160.
- ³ MITSCHERLICH, E. A., *Bodenkunde*, Berlin, 1913, p. 70.
- ⁴ MITSCHERLICH, E. A., *Bodenkunde*, Berlin, 1913, p. 30.
- ⁵ MUTTERICH; from E. RAMANN, *Bodenkunde*, Berlin, 1911, p. 398.
- ⁶ RAMBAUT, A. A., "Underground temperature at Oxford as determined by means of five platinum resistance thermometers." *Radcliffe Observations*, 1898-1910, p. 51.
- ⁷ GREENE, H., "Soil profile in the Eastern Gezira." *J. Agric. Sci.*, 1928, **18**, pp. 518-530.
- ⁸ HAINES, W. B., "The volume changes associated with variations of water content in soil." *J. Agric. Sci.*, 1923, **13**, pp. 296-310.
- ⁹ ATTERBERG, A., "Die Plastizität der Tone." *Int. Mitt. Bodenkunde*, 1911, **1**, pp. 10-43.
- ¹⁰ HARDY, F., "Cohesion in colloidal soils." *J. Agri. Sci.*, 1925, **15**, pp. 420-433.
- ¹¹ JOHANNSSON, S., "Die Konsistenzkurven der Mineralböden." *Int. Mitt. Bodenk.*, 1914, **4**, pp. 418-431.
- ¹² HAINES, W. B., "Studies in the physical properties of soils." I.—Mechanical properties concerned in cultivation. *J. Agric. Sci.*, 1925, **15**, pp. 178-200.
- ¹³ ANNETT, H. E., "The nature of the colour of black cotton soil." *Mem. Dept. Agr. India, Chem. Series*, 1910.
- ¹⁴ HUTTON, J. G., "Soil colours; Their nomenclature and description." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 164-172.

CHAPTER VIII.

WATER RELATIONSHIPS OF SOILS.

HYGROSCOPIC MOISTURE.

WHEN a soil is allowed to dry in an ordinary atmosphere, a certain amount of moisture is retained by hygroscopicity in the form of thin films of approximately molecular thickness. The percentage of hygroscopic moisture in a given soil varies with the state of saturation of the atmosphere with respect to water vapour. In equilibrium with a perfectly dry atmosphere, the hygroscopic moisture is completely lost, whilst in equilibrium with a saturated atmosphere the amount thus retained reaches a maximum. The relationships between moisture content and the degree of saturation of the atmosphere with which the soil is in equilibrium have been investigated by A. N. Puri, E. M. Crowther, and B. A. Keen¹, who obtained curves of a sigmoid type showing inflexion in the region of 50% humidity.

Since the atmosphere rarely attains complete saturation, the amount of hygroscopic moisture in air-dry soils is generally less than the maximum hygroscopic capacity. In order to determine the maximum hygroscopic capacity of a soil, it is allowed to attain equilibrium with a saturated atmosphere. Since the maximum amount of water vapour which can be retained by the atmosphere varies with the temperature, it is necessary, in making determinations of maximum hygroscopic capacity, to avoid fluctuations in temperature, for any cooling of the saturated atmosphere would lead to the deposition of dew on the soil. In practice, therefore, this determination is beset with considerable difficulty and

it is generally preferred to work with a standard atmosphere having a definite percentage saturation. On account of temperature effects and the slowness with which equilibrium with a saturated atmosphere is attained, it may be doubted if any of the values hitherto obtained represent full hygroscopic capacity.

The degree of saturation of a closed atmosphere can be controlled by using equilibrium solutions, generally sulphuric acid, of varying strength. Thus, in the Bureau of Soils, Washington, 99.8% saturation is obtained by using 2% (by vol.) sulphuric acid. Puri, Crowther, and Keen propose 50% saturation, attained in equilibrium with sulphuric acid of density 1.3321. It is shown that the value obtained in equilibrium with a 50% saturated atmosphere does not differ greatly from the moisture content of ordinary air-dry soil, since there is a fairly wide range on either side of this point over which the hygroscopic moisture content does not vary greatly. The significance of hygroscopicity determinations has been discussed by A. N. Puri².

However valuable for comparative purposes may be the hygroscopic capacity determined in equilibrium with a standard unsaturated atmosphere, it should be realised that the figures do not give the maximum amount of water which the soil can hold by hygroscopic attraction without being visibly moist.

The hygroscopic capacity as ordinarily determined on air-dry samples, although it gives an approximate relative measure of the full hygroscopic capacity, is influenced by changes in atmospheric humidity. The same sample of soil may give different values on different days if there is any marked change in atmospheric conditions. Yet, on considering a large number of results, it is evident that the highest figures are given by soils rich in clay or organic matter and by soils rich in sesquioxides, such as laterites.

The effect of these constituents on the maximum hygroscopic capacity may be seen from some data from E. W. Hilgard³ in Table XX.

TABLE XX. — HYGROSCOPIC CAPACITY OF CERTAIN SOILS (HILGARD).

(From "Soils," by permission of The Macmillan Co., Publishers.)

Soil or Clay	Clay %	Humus %	Ferric Oxide %	Hygroscopic Capacity %
Sandy loam ...	2.94	0.55	1.64	2.48
Pipe clay ...	74.65	0.00	0.15	2.09
Ferruginous clay ...	28.15	?	12.10	9.33
Ferruginous laterite	?	3.33	41.00	19.66
Muck soil (Peat) ...	?	66.10	?	21.00

B. A. Keen and J. R. H. Coutts⁴ examined the correlation of the moisture content of soils in equilibrium with an atmosphere of 50% humidity with other soil constants. Whilst a very high correlation was found between hygroscopicity and clay content, the correlation between hygroscopicity and ignition loss was less close. Since ignition loss is, except in heavy clay soils, mainly due to organic matter, this result might appear to imply that organic matter makes a small contribution to the hygroscopic capacity of the soil. Against this may be set the high values obtained for the hygroscopic moisture of peat soils such as that recorded in Table X. In some further studies on Natal soils, J. R. H. Coutts⁵ found a much higher correlation between moisture content at 50% humidity and ignition loss, but this was not surprising in view of the clay character of the soils used, for the loss on ignition in such cases is mainly due to water of constitution. Further work is desirable to establish the correlation which undoubtedly exists between hygroscopicity and organic matter content. With the more convenient methods now available for determining organic carbon, such an enquiry might be conveniently undertaken.

In a soil low in colloidal material, hygroscopic moisture is retained as films of approximately molecular thickness on the surface of the mineral particles, and the amount present under standard conditions is a measure of the total surface.

In ordinary soils, however, there is a certain amount of colloidal gel material and with this is associated a proportion of moisture termed *gel water*. Except in sandy soils, where there is a considerable content of gel-free particles, it may be supposed that the hygroscopic moisture is mainly gel water.

E. A. Mitscherlich,⁶ who has used hygroscopicity as a measure of the specific surface, has attempted to distinguish between the total surface, which is concerned with the absorption of hygroscopic water, and which is largely due to colloidal material, and the exterior surface of the mineral grains and soil crumbs. He has sought to obtain a measure for exterior surface by determining the amount of absorption from an atmosphere containing an organic vapour such as benzene, it being assumed that the benzene does not enter the interior micellar structures of the colloidal material.

A soil which contains no more moisture than is indicated by its maximum hygroscopic capacity is incapable of yielding water to plants or of permitting microbiological activity. Further, although certain changes have been observed after long storage of air-dry soils, the rate of chemical change is inappreciable compared with the rate in moist soils, i.e., soils which contain more moisture than their maximum hygroscopic capacity.

MOIST SOILS.

From the standpoint both of soil genesis and also of plant nutrition, the investigator is principally interested in moist soils. It is only in such soils that the pedogenic processes and the complex changes, involving the microbiological population and surface vegetation, take place.

The moisture conditions in the soil are governed to a large extent by climate, by the balance of rainfall and evaporation, and by the topographical situation of the soil, and also by the character of the horizons which compose the soil profile. In this connexion, the soil profile must be given a wide meaning, for, in many cases, the water conditions are controlled to a large extent by horizons lying below the

ordinary range of the soil-forming processes. It has been repeatedly emphasised that the unit of study is the soil profile, and it should be realised that studies of laboratory samples, though necessary for the elucidation of fundamental principles, can only give partial pictures of the conditions which obtain in nature.

It is a matter of common observation that even under similar climatic conditions, soils vary greatly in their water relationships. Generally speaking, sandy soils are dry and readily allow the rainfall to drain away from them, whilst clay soils are wet and retentive of the moisture obtained from the rainfall. It is important to understand the mechanism of water retention, and also to possess methods of comparing this property in different soils.

RETENTION OF WATER IN MOIST SOILS.

Much of the earlier thought on soil moisture was dominated by what has been termed the "capillary tube hypothesis." The soil moisture was considered to be present in the fine capillaries of the soil, which was pictured as behaving like a bundle of capillary tubes of varying diameter. Now $h = 2T/g\rho r$, where h = height of a water meniscus above ground water level; T = surface tension of water; g = gravitational constant; ρ = density of water; and r = radius of the capillary. Substituting the numerical values in this equation we obtain: $h = 0.15/r$ cm.

The large capillary rise implied by this equation has been called into question by numerous investigators, notably by B. A. Keen⁷, who has brought forward evidence (see p. 189) to show that, even in heavy loams, capillary rise is ineffective over distances of more than 80 cm. Although other writers have assigned a greater limit to capillary rise, it is clear from recent studies that the estimates of earlier workers for the height to which water can rise from a water table were far in excess of the truth. It is therefore evident that the conception of soil as a bundle of capillaries is inadequate to explain the observed facts.

Of recent years much attention has been devoted to working out the distribution of moisture in an *ideal soil** which, for this purpose, is pictured as consisting of uniform spheres in close (hexagonal) packing. At low moisture contents, the water is present as separate rings around the points of contact of the spheres. With increasing moisture content, the rings become connected with each other so as to form a continuous system interpenetrated by a continuous network of air spaces and passages.

Whilst conditions approximating to those in the "ideal soil" can scarcely ever be realised in actual soils, as a working hypothesis it marks a considerable advance on the simpler capillary tube hypothesis and has thrown considerable light on water distribution and movements.

In natural soils, the water relationships are dominated by the presence of colloidal gel material. The colloidal complex, which is partly inorganic and partly organic, has in common with other gels the property of retaining large proportions of water, termed *water of imbibition*. The water thus imbibed by the soil colloids is partly associated with the gels themselves, the *gel water*, and partly held in the interstices of the gel structure, the so-called *vesicular water*. It is only when the water content of a soil exceeds that represented by the imbibitional capacity of the colloidal gel material that external films, such as those contemplated in the ideal soil hypothesis, appear. In soils of low colloidal content, this point is reached when comparatively small proportions of moisture are present. The closest agreements of observed data with deductions from the theory of the ideal soil are, therefore, to be expected with sandy soils.

MOISTURE CAPACITY.

We may now consider the expression of the capacity of a soil to retain water. It should be remarked at the outset that there is some choice of methods of expressing the water

*The epithet ideal is, of course, used in a purely philosophical sense and implies no excellence for the growth of plants.

content of a soil. If the weight basis be chosen, the result may be expressed either on the moist soil or on the dry soil. For most purposes, the latter mode of expression is preferable, since it gives a truer picture of the amount of water associated with a unit weight of soil. For some purposes, it is convenient to express results as a volume percentage. This method is appropriate when results are to be considered in connexion with pore space. Here the results may be calculated to the volume of the total soil or to the volume of the actual soil particles. The relationship of these modes of expressing water content may be seen from the following example (Table XXI) for a hypothetical soil having apparent density 1.25, true density 2.50, and pore space 50%.

TABLE XXI. — CORRESPONDING PERCENTAGES OF WATER BY WEIGHT AND VOLUME FOR A TYPICAL SOIL.

Percentage moisture by weight		Percentage moisture by volume	
Moist soil	Dry soil	Total soil	Actual soil
28.6	40	50	100
24.4	32	40	80
19.4	24	30	60
13.8	16	20	40
7.4	8	10	20

The maximum water capacity is, of course, equal to the pore space, and is only realised in completely waterlogged soils. A measure of the water holding capacity of soils once widely used is the so-called *maximum water holding capacity*, obtained by determining the amount of water which a given quantity of soil can retain when placed in a shallow cylindrical vessel with a perforated bottom, saturated with water and allowed to drain freely. Various modifications of Hilgard's original method have been proposed, for example, that of B. A. Keen and H. Raczkowski⁸, but it should be noted that the results thus obtained approximate closely to the pore space, for, under such conditions, only the larger interstices of the soil will be void of water.

VERTICAL VARIATIONS IN MOISTURE CAPACITY.

The capacity of a soil to retain water cannot be defined without reference to the soil horizons which lie below it. Just as, from the standpoint of soil genesis, the unit of study is the soil profile, so in studies of soil moisture, the system to be considered is the whole succession of horizons down to the water table, or if this is very deep, down to the limit of atmospheric influences. And, therefore, although laboratory studies on soil moisture problems are indispensable for the elucidation of the fundamental principles, the final synthesis can only be made face to face with actual soil profiles.

If a hole or trench be dug in the soil in a valley bottom, a level is reached at which there is complete saturation. If the hole is sunk still further, water will stand in it up to this level of complete saturation, which is called the *water table*. The water table usually follows the contours of the land in a less accentuated manner. The depth below the surface at which it is encountered increases as the sides of a valley or hill are ascended. In the bottoms of valleys it approaches the surface of rivers and lakes. Complications are produced by impervious strata, so that in some cases the water table may be quite near the surface in elevated situations. This is sometimes called a "perched" water table.

Owing to the predominance of rainfall over evaporation during the winter months, and of evaporation over rainfall during the summer months, the position of the water table fluctuates, generally reaching a maximum at the end of the winter and a minimum at the end of the summer. Abnormal rainfalls or droughts may disturb this general succession.

In many soils, even in humid climates, the water table lies so far below the surface as to have practically no effect on moisture conditions at the surface. Thus, in the chalk upland soils of England, the water table is scarcely a factor in the soil moisture problem. It is difficult to assign a limit below which the water table does not affect conditions in the surface soil. It would be fairly safe to assume that, where

this level is more than twenty feet below the surface, the moisture relationships are mainly controlled by the conditions within the layer of free percolation. On the other hand, a water table lying within four feet of the surface has a definite effect on the moisture relationships at the surface. For example, a certain coarse sandy soil in Anglesey is in agricultural use in low-lying situations where the water table is within three feet of the surface, whilst in slightly elevated situations it is blowing dune.

The influence of position relative to a water table on water capacity is illustrated by data obtained by E. W. Hilgard⁹ in a sandy alluvium in California (Table XXII).

TABLE XXII.—VARIATION IN MOISTURE CONTENT IN A SANDY ALLUVIUM AT DIFFERENT DISTANCES FROM THE WATER TABLE (HILGARD).

(From "Soils," by permission of The Macmillan Co., Publishers.)

Inches above										
water table	1	6	12	18	24	30	36	42	47	
Moisture % ...	36.6	35.0	32.5	27.6	21.4	15.3	12.0	10.3	4.3	

The moisture content immediately above the water table will, of course, correspond closely with complete saturation of the pore space, as in the determination of maximum water holding capacity, and the gradual decrease in moisture content towards the surface represents the changing equilibrium between gravity and capillary forces.

MOISTURE EQUIVALENT.

When a soil is saturated up to its maximum water holding capacity a large proportion of the water is only loosely held in the capillary spaces. In the determination of *moisture equivalent* soil is placed in cups with perforated bottoms and submitted to a centrifugal force equivalent to 1000-2000 times that of gravity. The water held in the coarser interstices is thus removed and the only moisture retained is the imbibitional moisture of the colloidal gels and water in the smaller interstices or in thin external films.

A. F. Lebedev¹⁰ has shown that the value obtained for the moisture equivalent varies with the centrifugal force used and proposes the use of 18000 g., at which a limiting value, which he terms the *maximum molecular water capacity*, is reached.

Reverting to the data in Table XXII, if we assume that the moisture content at each level represents equilibrium, then any moisture in excess of the equilibrium moisture content for a given level will be subject to gravity and will find its way to the water table. Earlier writers have sometimes drawn a distinction between *gravitational water* and *capillary water*, gravitational water being any water in excess of that which the soil can hold by capillary attraction. It is evident that the limit between capillary water and gravitational water will vary according to the position with respect to the water table. Further, an increase in the force of gravity such as is realised in the determination of moisture equivalent will transfer a certain amount of capillary water to the category of gravitational water.

“STICKY POINT.”

F. Hardy¹¹ has proposed, as a measure of the water capacity of soils, the moisture content of a soil-water paste which can just be worked in the hands without being adhesive—the so-called *sticky point*. This is considered to represent the point at which the maximum imbibitional capacity of the soil colloidal material is satisfied. But since a sticky point can be obtained for fine sands and for kaolin, in which colloidal gel material is present in very small amounts, the moisture present at this limit must include some interstitial water.

B. A. Keen and J. R. H. Coutts⁴ have determined the sticky point of a number of soils from which most of the colloidal organic matter had been removed by treatment with hydrogen peroxide. The relationship with the clay content is approximately linear. Plotting the “sticky point” moisture, Sp, against the clay, the curve cuts the Sp axis at

about 16% moisture. From this it may be deduced that a soil with no colloidal material should show a sticky point of approximately 16% moisture by weight. For an "ideal soil" with close packing the moisture content calculated on a weight basis with the interstices filled would be 14.6%

B. de C. Marchand¹² finds, for Cape Western Province soils, that the sticky point water, expressed as a volume percentage, is given by the equation

$$S_v = 0.58 \text{ Clay} + 26 \pm 1.2$$

The value 26 in this equation is in complete agreement with the theoretical pore space for an ideal soil—a result which can only be possible by reason of compensating adjustments, since the actual particles of soils are neither spherical nor of uniform size.

The contribution of organic matter to the "sticky point" water capacity of soils has not yet been determined, but it is known that highly organic soils show high "sticky points."

It should be added that the moisture conditions in a soil at the "sticky point," as determined in the laboratory, are rarely, if ever, attained under natural conditions. In this determination, the moistened soil is brought into the single-grain or puddled condition and the paste consists simply of soil with water completely saturating the colloidal gels and filling the interstices.

IMBIBITIONAL MOISTURE.

A. F. Joseph¹³ has determined the imbibitional moisture in the clay fractions of a series of soils by obtaining for each soil the moisture equivalent and the xylene equivalent. Assuming that, in the case of xylene, there is no imbibition by gels, then the difference between the two values should give the imbibitional moisture, i.e., the moisture associated with the colloidal gel material.

Some results obtained by this method are given in Table XXIII.

TABLE XXIII.—MOISTURE EQUIVALENT, XYLENE EQUIVALENT, AND IMBIBITIONAL MOISTURE OF CLAY FRACTIONS WITH DIFFERENT CATIONS (JOSEPH).

	Original Clay	Sodium Clay	Calcium Clay
Moisture equivalent by weight ...	68.7	125.0	63.0
Xylene equivalent by weight ...	11.7	11.4	11.1
Imbibitional moisture by volume	143.0	292.0	131.0

Some interesting points arise from these data. In the first place, it will be noticed that the xylene equivalent is practically the same for the original clay, the sodium clay, and the calcium clay. The imbibitional moisture is, however, far higher in the sodium clay than in the other clays. This is in agreement with the highly dispersed character of sodium soils. The original soil, containing principally calcium as its exchangeable base, does not markedly differ from the calcium soil. The clay fractions of alkaline soils would probably approximate to sodium clays.

In Table XXIV are given a selection from Joseph's data for the imbibitional moisture of some representative clays.

TABLE XXIV.—IMBIBITIONAL MOISTURE AND SILICA-ALUMINA RATIO FOR CLAY FRACTIONS (JOSEPH).

	$\text{SiO}_2/\text{Al}_2\text{O}_3$	Imbibitional moisture Volume %
China clay ...	2.11	5.0
Ball clay ...	2.66	39.5
Subsoil clay, N. Wales ...	2.84	101.6
Kassala alluvium ...	3.77	79.2
Badob ...	3.90	100.8
Blue Nile alluvium ...	4.53	128.0
Bentonite ...	6.96	261.0

RELATIONSHIPS BETWEEN SOIL MOISTURE CONSTANTS.

It is evident that the different constants which have been proposed as measures of the water holding capacities of soils, since they depend on texture and colloidal character, must be connected among themselves. Much ingenuity has therefore been shown in devising numerical relationships among them.

L. J. Briggs and H. L. Shantz¹⁴ suggested an empirical formula connecting the maximum water holding capacity (M) of a soil with its hygroscopic capacity (H), namely,

$$M = 4.3H + 21$$

The term $4.3H$ in this formula may be equated with the "sticky point" moisture of Hardy.

With spherical particles in close packing, the water retained as films was calculated by B. H. Wilsdon¹⁵ to be 23.46%, which is considered to be represented by 21% in the Briggs-Shantz equation. Wilsdon connects the colloidal moisture capacity (Hardy's "sticky point") with the hygroscopic capacity by the formula

$$P = aH$$

and finds a value $a = 4.73$ for an Indian soil. Hardy found values ranging from 2.73 to 4.68 with West Indian soils, the lowest figures being obtained with laterites and ferruginous soils. In Wilsdon's equation, $3.73H$ represents the vesicular water, whilst H represents the gel water, equated with the hygroscopic capacity.

L. J. Briggs and J. W. McLane¹⁶ found a general relationship connecting the moisture equivalent (ME) with the hygroscopic co-efficient (H), $ME = 2.7H$. J. C. Russell and W. W. Burr¹⁷ similarly found $ME = 2.24H$. From these relationships it appears that, at the point represented by the moisture equivalent, a certain amount of the vesicular water must have been removed.

WATER MOVEMENTS IN SOILS.

The capillary tube hypothesis has been used by many workers in developing the theory of water movement in soils.

W. H. Green and G. A. Ampt¹⁸ based their treatment of the problem on Poiseuille's equation for the flow of a liquid through a cylindrical tube :—

$$v = \pi g h \rho r^4 / 8 \eta l,$$

where v = vol. of liquid passing in time t , g = gravitational constant, h = head of pressure, ρ = density of liquid, r = radius of tube, η = viscosity of liquid and l = length of tube.

In the downward percolation of water through a saturated soil column of length l under a head of water a , the above equation becomes :—

$$v = \pi g (a + l) \rho \Sigma r^4 / 8 \eta l,$$

in which r is the effective average radius of the capillaries. Defining permeability (P_w) as the volume of liquid flowing in unit time through a column of unit length and cross section under unit head of pressure, we have

$$P_w = (\pi g / 8 \eta) \cdot (\Sigma r^4 / A),$$

where A = actual cross section of column ; whence

$$v = P_w t A (a + l) l.$$

Equations have been worked out for downward percolation of water, upward motion, and horizontal motion, in which terms are introduced for the action of capillary forces. With downward movement, the velocity decreases hyperbolically from infinity to a constant value P_w/S , where S is the pore space. In upward movement, the velocity decreases hyperbolically from infinity to zero, when the term denoting the capillary pull is equal to the height of the rising water front. In horizontal motion, the velocity is inversely proportional to the distance travelled by the water front and also decreases hyperbolically to zero.

Experimental data with coarse materials were found to fit fairly well with the above deductions, but it became evident that the problem had been over-simplified by the assumption of complete saturation behind the water front, and certain modifications were introduced. More serious divergences, however, were introduced in clay soils by the swelling of colloidal material on wetting.

The problem of water movement has also been attacked by introducing the conception of capillary potential, analogous to temperature gradient in the theory of conduction of heat. A discussion of this direction of enquiry is, however, outside the scope of the present book and the student is referred to the papers of W. Gardner¹⁹, E. Buckingham²⁰, and B. H. Wilsdon¹⁵.

More recently the study of the movements of water in soils has been affected by the "ideal soil" hypothesis, and the student may be referred to the discussions of W. B. Haines²¹ and R. A. Fisher²².

DISTRIBUTION OF ADDED WATER IN THE SOIL PROFILE.

Great ingenuity has been shown in developing the theory of the distribution and movement of water in ideal soils under laboratory conditions. But it must be realised that, in the majority of soils encountered in the field, the situation is dominated by the presence of colloidal gel material.

In very wet soils, a considerable proportion of the moisture present is free to move downwards under the influence of gravity and, eventually, an equilibrium distribution is attained depending on the position of the water table and the colloidal moisture capacity of the different horizons of the profile.

We may consider first the case of a soil profile in which the water table lies so far below the surface as to be without effect on the moisture distribution in the upper horizons. An excess of water applied to the surface horizons will distribute itself so as to attain capillary equilibrium down to the water table. In the absence of evaporation, assuming uniform water holding capacity, the moisture content down to the vicinity of the water table will approximate to what is termed by J. A. Widtsoe and W. W. McLaughlin²³ the *lento-capillary point*, i.e., that moisture content at which capillary movements become sluggish and ineffective.

If water is added by rainfall or irrigation to the surface of a soil in which the moisture content throughout the profile is less than that corresponding with the lento-capillary point, it will tend to distribute itself in such a way that the soil is moistened up to this point, so far as is possible, with the water available.

The distribution of moisture applied to the surface of a dry soil is shown by some experiments by C. F. Shaw²⁴. Using 36 inches of dry soil and applying a 6 inch head of water it was found that after 24 days the moisture content was fairly constant at about 20% down to 19 inches and then fell off to 3.33% at about 30 inches. After 123 days the moisture content decreased slightly from 17.32% at the surface to 16.16% at 24 inches. It was concluded that this represented practically the equilibrium distribution of the water applied, apart from possible movements in the vapour phase. The average value, 16.79%, termed by Shaw the *normal moisture capacity*, approximates fairly closely to the moisture equivalent for the same soil, namely, 15.88%. These results are shown graphically in Fig. 9. It is suggested that, for moisture contents up to the normal moisture capacity, capillary movement is negligible. The normal moisture capacity of Shaw is thus apparently the lento-capillary point of Widtsoe and McLaughlin, but appears to lie above rather than below the moisture equivalent. The subsequent drying out of the soil would take place entirely by evaporation, which would not necessarily be entirely at the surface, but might involve vapour movements from lower levels. In this connexion, the influx and efflux of air consequent on diurnal temperature variations is probably of importance.

Shaw's results were obtained with a uniform soil under laboratory conditions. In a natural profile, the variations in the nature and content of colloidal material throughout the profile would result in a different distribution of moisture. Presumably the presence of a considerable proportion of organic matter in the surface soil would be reflected in a

higher equilibrium moisture content. The march of the curve down to the limit of wetting would, indeed, tend to follow that of the moisture equivalent.

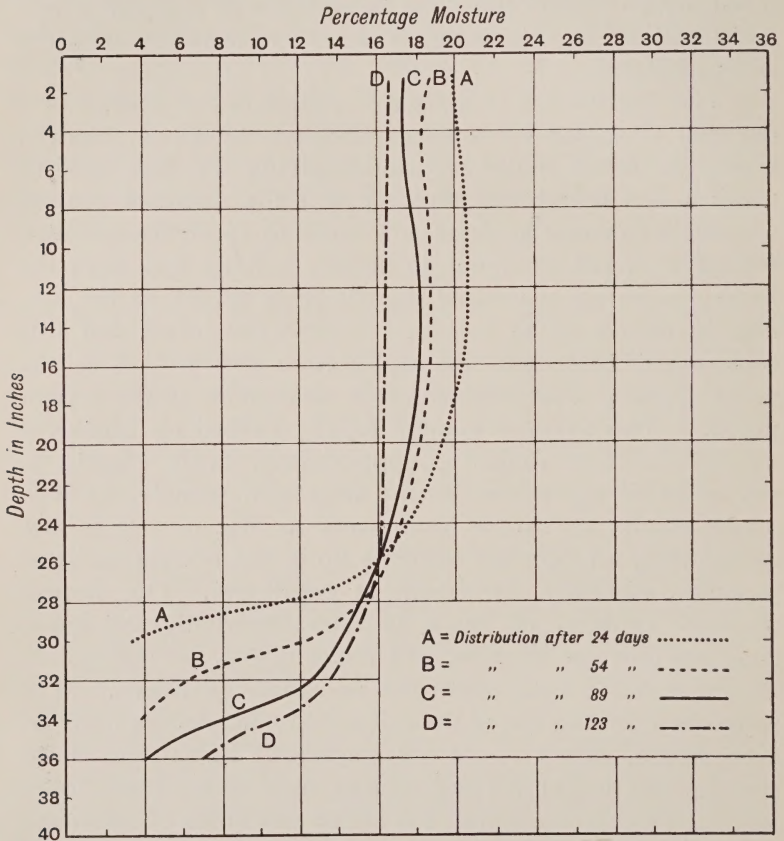


FIG. 9.—Distribution of water in tubes receiving a 6-inch irrigation, after prolonged periods of adjustment. (C. F. SHAW.)
 (From "Soil Science"—by permission.)

F. J. Veihmeyer and A. H. Hendrickson²⁵ have studied the *field capacity* of certain irrigated soils in California. By field capacity is understood the "amount of water held in the soil after excess gravitational water has drained away and the rate of downward movement has materially decreased."

This, in practice, is equivalent to the "normal field capacity" of Shaw. For all except the lightest soils, field capacity approximates closely to the moisture equivalent as determined by the modification of the method of L. J. Briggs and H. L. Shantz, the average ratio being 1.06 ± 0.023 . In the absence of evaporation, from two to three days after irrigation are required for the field capacity to be reached, but conditions are materially altered by the presence of impervious layers. If water is applied by irrigation to a dry soil it appears that, when equilibrium is reached, there is a fairly sharp transition from moistened to unmoistened soil, all the moist soil showing approximately the same degree of wetness relative to the moisture equivalent. This is readily intelligible when it is realised that capillary transmission becomes negligible at the lento-capillary point.

It is perhaps necessary to draw a distinction between the apparently interchangeable terms "lento-capillary point" and "normal moisture capacity." Whilst the former term applies to the isolated soil and is a moisture constant determinable under laboratory conditions, the "normal moisture capacity" relates to field conditions. And thus, whilst the two terms may be interchangeable in referring to a deep profile unaffected by a water table, in the case where a water table is near the surface, the "normal moisture capacity" will be considerably greater than the moisture content at the "lento-capillary point." It would appear that the lento-capillary point approximates to the moisture equivalent.

In the presence of a water table near the surface, the equilibrium moisture distribution is modified. In the Shaw experiment, a water table at, say, 24 inches, below the surface would be reflected in an increased moisture content, as in the Hilgard experiment mentioned on page 180. The presence of impervious or relatively impervious strata would have the same effect as a water table.

In an experiment described by B. A. Keen⁷, deep cylinders 16 feet by 2 feet were sunk in the ground with their top

edges slightly above the surface. They were then filled with different kinds of soil. By means of a narrow tube in each cylinder it was possible to determine the position of the ground-water level. The data obtained show how the water table in an initially saturated soil falls during drought. Since the fall in level for coarse sand practically ceases when the water table is 35 cm. below the surface, it is clear that capillary rise has ceased to be effective in transmitting water to the surface. The corresponding values for fine sand and heavy loam were 70 cm. and 80 cm., respectively. It may, therefore, be assumed that these figures represent the depths below which the effect of a water table on moisture conditions at the surface may be regarded as inappreciable.

It would be of considerable interest to determine, for some typical profiles, the moisture distribution attained after free drainage in the absence of losses by evaporation or gains from rainfall. Under such circumstances the moisture content would not fall below that represented by the field capacity or normal moisture capacity, and this would represent the lowest moisture content to which a soil could be reduced by drainage alone. For a moderately heavy clay, this would imply a moisture content in the region of 30% as the limit, in the absence of removal by evaporation. It is a matter of common knowledge that, under British conditions, clay soils are persistently wet in winter, even although little rain falls. The reduction of the moisture content of such soils to the point at which they can be worked is mainly dependent on the occurrence of drying winds, and cannot be effected by drainage alone. Similar considerations apply to highly organic soils such as those of parts of Wales.

LOSSES OF MOISTURE BY EVAPORATION.

Under a covering of natural vegetation, losses from the soil to the atmosphere are principally by transpiration from the leaves of plants. Under crops, the losses are partly by evaporation from bare soil. The rate of loss by transpiration and evaporation depends partly on meteorological factors

such as temperature, humidity, and wind velocity, and partly on the character of the transpiring vegetation or evaporating soil. Other things being equal, moisture losses are less under a covering of vegetation than with bare soil.

The principal circumstance affecting the evaporation of moisture from soils is the temperature. This is illustrated by the data, reported by A. D. Hall²⁶, obtained from the Rothamsted lysimeters. The amount of percolation is measured from columns of soil in which the original structure has been preserved unaltered. The surface of the soil is maintained tilled but uncropped. The rainfall is completely accounted for by evaporation and percolation since there is no run-off from the surface. Hence, by measuring the percolation, the amount of evaporation can be determined. The data are shown in Table XXV.

TABLE XXV.—TEMPERATURE, RAINFALL, PERCOLATION, AND EVAPORATION AT ROTHAMSTED (HALL).

Month	Rainfall Average	Temperature 26 years average	Percolation through 60 inch gauge	Evaporation	Daily Evaporation
		deg. Fahr.	inches	inches	inches
Jan. ...	2.32	36.6	1.96	0.36	0.0116
Feb. ...	1.97	38.2	1.48	0.49	0.0175
Mar. ...	1.83	40.9	0.95	0.88	0.0284
Apr. ...	1.89	45.5	0.53	1.36	0.0453
May ...	2.11	51.2	0.50	1.61	0.0520
June ...	2.36	57.5	0.62	1.74	0.0580
July ...	2.73	60.7	0.65	2.08	0.0671
Aug. ...	2.67	59.9	0.58	2.09	0.0674
Sept....	2.52	55.9	0.76	1.76	0.0587
Oct. ...	3.20	48.0	1.68	1.52	0.0490
Nov....	2.86	42.6	2.04	0.82	0.0273
Dec. ...	2.52	37.7	2.04	0.48	0.0156

It will be seen that there is a close correlation between temperature and the daily evaporation, and it can be readily understood that, in tropical climates, the rate is still greater.

Where losses are due to transpiration from plants, the temperature effect will be similar.

Evaporation is markedly increased by wind, particularly if the atmospheric humidity is low. The rapid drying out of soils in east winds is well known to practical agriculturists.

The effect of variations in atmospheric humidity is shown by some data from Masure, cited by E. Ramann²⁷ (Table XXVI).

TABLE XXVI.—EFFECT OF VARYING HUMIDITY ON RATE OF EVAPORATION FROM SOIL (MASURE).

Temperature °C	Relative Humidity %	Evaporation mm.
17.6	74	0.93
17.7	79	0.62
17.0	89	0.38
17.2	91	0.25

The great differences in the rate of evaporation depending on temperature, humidity, and wind velocity, involve corresponding variations in the amount of percolation through the soil. And therefore the rainfall, in itself, is a very inadequate measure of the leaching to which a soil is subjected. With a mean annual temperature of 5°C, varying between 0° and 10°C, a mean annual rainfall of 20 inches would be sufficient to furnish drainage. The same rainfall, with a mean annual temperature of 25°C, varying between 15°C and 30°C, would be insufficient to outweigh evaporation.

The rate of evaporation from soil is governed not only by environmental factors but also by soil conditions. The principal factor is the moisture content. Generally speaking, in laboratory experiments with thin layers of soil, the rate of evaporation is constant over a wide range of moisture content, but tends to fall off when the region of the hygroscopic co-efficient is approached. Under field conditions, the problem is more complex, for the rise of capillary water to the surface plays a part. But when the moisture

content at the surface is reduced to the lento-capillary point, no further replacement from below is possible and further losses are by evaporation from the body of the soil down to the level at which capillary movement is still possible.

The successive effects of irrigation and evaporation are well shown by some observations of H. Greene²⁸ in the

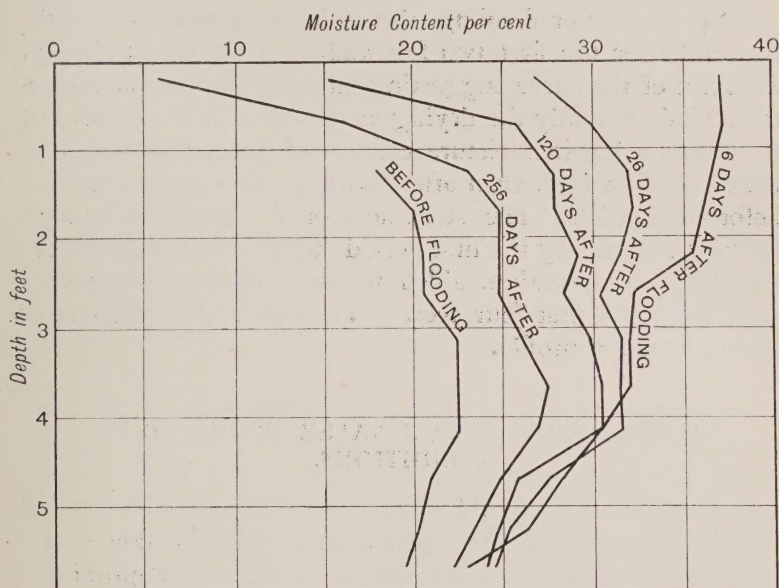


FIG. 10.—Drying out of land after 14 days' flooding. (H. GREENE.)
(From "Journal of Agricultural Science"—by permission.)

Sudan Gezira. The soil used was a heavy æolian clay having the following soil moisture constants:—

Moisture equivalent about	...	...	40%
Lento-capillary point about	...	...	28%
Hygroscopic co-efficient about	...	...	14%

It will be noticed that the lento-capillary point for this soil lies considerably below the moisture equivalent.

The results of the investigation are shown diagrammatically in Fig. 10. Before irrigation, the moisture con-

tent of the soil varied about 20% down to 6 feet. Assuming the absence of capillary movement below the lento-capillary point, this implies that evaporation had taken place not only at the surface, but down to the level explored. The tendency for the curve connecting the moisture content of the unflooded soil to swing to the left at the top reflects the more intense evaporation at the surface.

Six days after flooding, the moisture content was fairly constant over the first two feet and then fell off gradually, the form of the curve suggesting an effect down to at least 6 feet. Subsequently the drying at the surface was rapid, but there was a lag in moisture content at lower levels so that below the first foot, even after 256 days, it was greater than before irrigation. The steadiness of the moisture content below 4 feet during the first period suggests that losses were mainly by evaporation, aided to some extent by capillary rise, from the first four feet. Thereafter, losses occurred throughout the profile.

LOSSES AND GAINS OF WATER UNDER FIELD CONDITIONS.

Movements of liquid water in soils are mainly due to gravitational water. The penetration of water downwards into dry soil depends on the saturation or approximate saturation of superficial layers. After the soil has been dried out to a considerable depth during summer drought, the penetration of moisture is very slow and, under British conditions, the sub-soil sometimes remains fairly dry until mid-winter.

The range of upward capillary movement is now known to be comparatively small. In Keen's experiments, 80 cm. appeared to be the limit of upward movement. Lateral capillary movement of moisture is also confined to small distances. In the absence of water percolating down slopes, soil under cover remains dry, even if surrounded by soil almost completely saturated with moisture.

Under field conditions, the absorption and disposal of rainfall depends on the texture and structure of the soil. Sandy soils rapidly absorb rainfall and allow ready percolation to the water table. Clay soils oppose a greater resistance to water movements, but differences are observable depending on structure and on the occurrence of compacted horizons. Lateritic clays are more open in structure and thus less subject to surface washing under heavy rainfall than aluminosilicic clays, which may not be able to dispose of intense rainfall sufficiently rapidly to prevent run-off and consequent erosion.

Owing to continued interference through gains from rainfall on the one hand and losses by evaporation on the other, the attainment of moisture equilibrium in a soil profile must be regarded as a purely ideal condition. It is of interest, however, to consider the type of moisture distribution which might be expected in a soil profile with a water table at such a depth as to be without influence on surface conditions.

During a period when rainfall predominates over evaporation, the tendency will be towards a distribution in which the moisture content at each level will approximate to that represented by the lento-capillary point or field capacity. This does not necessarily imply an uniform distribution, since the lento-capillary point will vary with the colloidal content and also, to some extent, with the structure. Uniform distribution would only be approached in an uniform soil. Excess of moisture over this ideal distribution, owing to recent precipitation, would be reflected in a higher moisture content at the surface. In the absence of evaporation this would be translated to lower levels partly by gravitational movement and partly by capillary transmission and would increase the depth of soil moistened to the lento-capillary point. If the depth thus moistened extends to the zone affected by the water table, any excess of moisture would be handed on to the ground-water. The rapidity with which such an adjustment is attained will depend on the

permeability of the soil to water movements. In heavy soils, adjustments are slow and moisture contents considerably above the equilibrium distribution may persist in the surface horizons for long periods.

With soils of low permeability, heavy rain may result in the saturation of the surface soil. Further precipitation is then disposed of by superficial run-off, with the possibility of erosion of the top soil. The susceptibility of soils to erosion in this way depends on their texture, structure, organic matter content, and the composition of the clay complex. Generally speaking, the tendency to a deflocculated or single-grain structure implies liability to erosion by run-off.

We may now consider the changes in moisture distribution consequent on the cessation of precipitation and the onset of conditions favouring evaporation. If the surface soil is saturated above its equilibrium moisture content by recent rain, losses will take place both by evaporation and also by capillary and gravitational translation to lower levels. Losses by movement to lower levels will cease when the equilibrium moisture distribution has been reached, and thereafter losses will be entirely by evaporation, at first mainly from the surface, but subsequently, as in Greene's experiments, from the lower horizons. In the earlier stages of drying out there is also the possibility of upward capillary movements, but as each successive layer becomes dried to the lento-capillary point, the range of upward movement will be restricted and the losses by evaporation from the superficial horizons will no longer be replaced from below. Since the rate of direct evaporation is greatest at the surface and decreases with depth, the rate of loss of moisture from the soil as a whole must, *ceteris paribus*, decrease when capillary movement to the surface is no longer possible, and will be determined mainly by the direct evaporation from the surface and the subjacent layers.

Where a water table is present near the surface, conditions will be different, for capillary rise will suffice to replace the losses by evaporation from the surface. But from the

experiments of Keen it appears that, with a water table below 35 cm. for coarse sand, 70 cm. for fine sand, and 80 cm. for heavy loam, capillary rise can no longer cover the losses by evaporation from the surface layers.

It has been suggested by A. F. Lebedev²⁹ that, apart from movements of water in the fluid state, a certain amount of vertical translocation takes place in the vapour state. During the winter months, the surface soil is colder than the subsoil, and this may result in an upward distillation conformable with the difference in vapour pressure. Accession of moisture in this way may be a contributory cause of the persistent wetness of soils in late winter when the temperature gradient to the surface is steepest. On the other hand, during the summer months, a downward distillation may take place from the warm surface soil to the colder subsoil, thus reinforcing the losses by evaporation from the surface.

BIBLIOGRAPHY.

- ¹ PURI, A. N., CROWTHER, E. M., and KEEN, B. A., "The relation between the vapour pressure and water content of soils." *J. Agric. Sci.*, 1925, **15**, pp. 68-88.
- ² PURI, A. N., "A critical study of the hygroscopic coefficient of soil." *J. Agric. Sci.*, 1925, **15**, pp. 272-283.
- ³ HILGARD, E. W., *Soils*. New York, 1912, p. 196.
- ⁴ KEEN, B. A., and COUTTS, J. R. H., "'Single Value' soil properties: A study of the significance of certain soil constants." *J. Agric. Sci.*, 1928, **18**, pp. 740-765.
- ⁵ COUTTS, J. R. H., "'Single Value' soil properties: A study of the significance of certain soil constants." II.—Studies on Natal soils. *J. Agric. Sci.*, 1929, **19**, pp. 325-341.
- ⁶ MITSCHERLICH, E. A., *Bodenkunde*, Berlin, 1913, p. 58 *et seq.*
- ⁷ KEEN, B. A., "The limited role of capillarity in supplying water to plant roots." *Proc. Int. Congr. Soil Sci.*, 1928, Vol. I, pp. 504-511.
- ⁸ KEEN, B. A., and RACZKOWSKI, H., "The relation between the clay content and certain physical properties of a soil." *J. Agric. Sci.*, 1921, **11**, pp. 441-449.
- ⁹ HILGARD, E. W., *Soils*. New York, 1912, p. 208.
- ¹⁰ LEBEDEV, A. F., "Methods of determining the maximum molecular holding capacity of soils." *Proc. Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 551-563.
- ¹¹ HARDY, F., "The physical significance of the shrinkage coefficient of clays and soils." *J. Agric. Sci.*, 1923, **13**, pp. 243-264.
- ¹² MARCHAND, B. DE C., "The sticky point water of soils." *S. African Journ. Sci.*, 1930, **28**, pp. 183-193.

- ¹³ JOSEPH, A. F., "The moisture equivalent of heavy soils," II. *J. Agric. Sci.*, 1927, **17**, pp. 12-20.
- ¹⁴ BRIGGS, L. J., and SHANTZ, H. L., "The wilting coefficient for different plants and its indirect determination." *U.S. Dept. Agr. Bur. Plant Int.*, 1912, Bull. 230.
- ¹⁵ WILSDON, B. H., "Studies in soil moisture," Part I. *Mem. Dept. Agr. India, Chem. Series*, 1921, **6**, pp. 154-186.
- ¹⁶ BRIGGS, L. J., and McLANE, J. W., "The moisture equivalent of soils." *U.S. Bureau, Soils*, 1907, Bull. 45.
- ¹⁷ RUSSELL, J. C., and BURR, W. W., *Soil Sci.*, 1925, **19**, pp. 251-266.
- ¹⁸ GREEN, W. H., and AMPT, G. A., "Studies in soil physics." I.—The flow of air and water through soils. *J. Agric. Sci.*, 1911, pp. 1-24, *ibid.*, 1912, **5**, pp. 1-26.
- ¹⁹ GARDNER, W., "The movement of moisture in soil by capillarity." *Soil Sci.*, 1919, **7**, pp. 313-317. "Capillary moisture holding capacity," *ibid.*, pp. 319-324. "A capillary transmission constant, and methods of determining it experimentally," *ibid.*, 1920, **10**, pp. 103-126. "The capillary potential of the soil and its relation to soil moisture constants," *ibid.*, 1920, **10**, pp. 357-358. (With J. A. WIDTSOE.) "The movement of soil moisture," *ibid.*, 1921, **11**, pp. 215-232.
- ²⁰ BUCKINGHAM, E., "Studies on the movement of soil moisture." *U.S. Bureau Soils*, 1907, Bull. 38.
- ²¹ HAINES, W. B., "Studies in the physical properties of soils." II.—A note on the cohesion developed by capillary forces in the ideal soil. *J. Agric. Sci.*, 1925, **15**, pp. 529-535. IV.—A further contribution to the theory of capillary phenomena in soil, *ibid.*, 1927, **17**, pp. 264-290. V.—The hysteresis effect in capillary properties and the modes of moisture distribution associated herewith, *ibid.*, 1930, **20**, pp. 97-116.
- ²² FISHER, R. A., "On the capillary forces in an ideal soil." *J. Agric. Sci.*, 1926, **16**, pp. 492-505. "Further note on the capillary forces in an ideal soil," *ibid.*, 1928, **18**, pp. 406-410.
- ²³ WIDTSOE, J. A., and McLAUGHLIN, W. W., "The movement of water in irrigated soils." *Utah Exp. Sta.*, 1912, Bull. 115.
- ²⁴ SHAW, C. F. "The normal moisture capacity of soils." *Soil Sci.*, 1927, **23**, pp. 303-317.
- ²⁵ VEIHMEYER, F. J., and HENDRICKSON, A. H., "The moisture equivalent as a measure of the field capacity of soils." *Soil Sci.*, 1931, **32**, pp. 181-194.
- ²⁶ HALL, A. D., *The Book of the Rothamsted Experiments*, London, 1905.
- ²⁷ RAMANN, E., *Bodenkunde*, Berlin, 1911, p. 351.
- ²⁸ GREENE, H., "Soil permeability in the Eastern Gezira." *J. Agric. Sci.*, 1928, **18**, pp. 531-543.
- ²⁹ LEBEDEV, A. F., "Die Bewegung des Wassers im Boden und im Untergrund." *Z. Pflanz. Düng.*, 1927, **10A**, pp. 1-36.

CHAPTER IX.

SOILS OF THE PODSOLIC GROUP.

THE soils to be described in this chapter occur principally under humid conditions in cold to temperate regions. The podsols and their congeners are completely leached soils in which calcium carbonate and calcium sulphate are present only as fugitive constituents. The reaction is thus on the acid side of neutrality. The natural vegetation is generally forest or heath, but prairie vegetation is associated with one of the types, whilst swamp, fen, and marsh vegetation appear as local associations where drainage conditions are sufficiently impeded.

TUNDRA.

These soils occur in sub-arctic regions such as Northern Europe and Asia, and Northern Canada. They lie outside the limits of forest and are occupied by mosses, carices, dwarf willows, festucas, etc. In the tundra region, the dominant factors in soil formation are physical weathering and waterlogging. The latter is occasioned by the scanty evaporation, which results in the prevalence of a humid climate even where the rainfall is comparatively small. Chemical weathering and microbiological activity are naturally less pronounced than in warmer climates, but are not entirely suppressed. These soils are commonly assigned to the podsolic group, but the prevalence of waterlogging among the conditions of their development suggests that they would be more appropriately grouped with the ground-water soils.

The characteristic feature of the tundra profile is the occurrence, below a certain depth, of a perpetually frozen

layer, whose presence is, in some districts, not inconsistent with a fair level of fertility during the growing season. .

A typical tundra profile has been described by Ssukatshev¹. The succession is as follows :—

- I. Humous grey-brown horizon. Thickness about 3 cm., containing occasional undecomposed plant remains.
- II. Yellowish-grey horizon; occasionally greyish-brown, ochreous friable loam. Thickness 3-5 cm.
- III. Greyish-blue uniform sticky loam of semi-fluid consistency. Thickness 8-10 cm.
- IV. Brownish-yellow loam similar to horizon II., but more compact. Thickness 2-3 cm.
- V. A compact brownish-grey solid loamy horizon, often showing at 40-60 cm. dark fluid mottlings apparently of a humous character. Frozen below 79 cm. from surface, but retaining the same character for a further 10 cm.—the limit of exploration.

The bluish-grey horizon is apparently the characteristic mark of tundra, becoming thicker with increase and thinner with decrease of moisture. In sandy profiles, it disappears, presumably owing to better aeration.

When the surface of tundra becomes frozen, the semi-fluid greyish-blue horizon is held under pressure, owing to the expansion of water below 4°C, and this leads, in some cases to an eruption through the thin frozen layer, in other cases, to the surface becoming thrown into hummocks. The eruption of material leads to a mixing of the profile and tends to obliteration of horizons.

Tundra soils result mainly from physical weathering and are consequently poor in clay. They exhibit a natural tendency to peat formation. The peat horizon is often dissected by running water into a characteristic hummocky formation.

The occurrence of a perpetually frozen subsoil in the tundra region has been discussed by C. Nikiforov². Whilst the prevalence of a mean annual temperature below freezing point might sufficiently account for the existence of a permanently frozen layer, it has been held by some investigators that this phenomenon is a survival from the Glacial Period.

It has been stated that perpetually frozen subsoils occur outside the tundra region and may sometimes underlie the more northern podsoils, black earths, and chestnut earths.

PODSOLS.

This group of soils, by reason of its wide extent in Russia and Northern Europe, has received more detailed attention than any other group of soils, with the possible exception of the black earths.

The podsoils of the northern hemisphere occupy the humid regions lying to the south of the tundra: with their related types, they occupy the greater part of northern and western Europe. They are developed typically under coniferous forest, but also occur under heath vegetation which, in its meagre demands on the plant nutrients of the soil, exhibits a close resemblance to coniferous forest.

A typical podsol profile consists essentially of three horizons. The A, or eluviated horizon, consists of a layer of peaty material underlain by a more or less bleached layer, relatively poor in humus and sesquioxides. Below the A horizon is the B horizon. This is enriched by certain of the constituents leached from the A horizon. The constituents which distinguish this horizon by their accumulation are humus and sesquioxides. As the conditions of deposition vary with the relative proportion of humus and sesquioxides in the percolating moisture and also with its reaction, different kinds of podsol may be distinguished. In the more extreme types, the humus podsoils, the B horizon is enriched by an accumulation of humus and sesquioxides, the accumulation of sesquioxides occurring at a lower level in the profile than that of the humus. In less extreme podsoils, there may be no accumulation of humus, but an accumulation of hydrated ferric oxide in the B horizon. These are known as iron podsoils. The C horizon is the parent material from which the profile is developed. The precipitation of materials in the B horizon of podsoils often leads to the formation of hardpan (ortstein).

The dominant factor in the development of the podsol profile is the prevalence of intense leaching owing to the continued excess of rainfall over evaporation.

Whilst podsoles may be developed on all kinds of parent materials, the degree of development is affected by the base status of the substrate and by its permeability to leaching. The most strongly developed profiles occur in light sands which are naturally poor in basic constituents. In such cases, even where the climate is not of the most extremely humid type, impoverishment rapidly occurs and the only possible type of vegetation consists of plants such as conifers and heath plants which can exist on minimal supplies of mineral plant food. With poverty in basic constituents, the residues of vegetation undergo a type of decomposition, chiefly through the action of fungi, resulting in the production of a layer of acid peaty organic matter which, in the absence of earth worms, remains sharply differentiated from the mineral soil over which it lies. The extremely acid leaching results in the development of a deep bleached layer and a humous B horizon. Such podsoles may be seen on the Bagshot Sands of S.E. England. They are widespread on the sandy soils of Northern and Middle Europe.

Podsoles are less readily developed from materials with better base reserves than the light quartzose sands. On loams and clays the podsolised layer is shallower and indeed, in some cases, cannot be recognised without chemical analysis of the colloidal material. It is obvious, then, that the distribution of podsoles is markedly influenced by geology. If the whole of England were a region of light sands, such as the Bagshot Beds, it would be represented on a soil map as an area of podsoles. Actually, owing to the prevalence of clays and loams having notable base reserves, well developed podsoles are by no means common. It is possible that in certain upland regions, original podsoles have been truncated by erosion consequent on deforestation in former centuries. In such areas, secondary podsoles may be developed in the original B horizon, which has been laid bare by erosion.

It is unfortunate that the available data for podsol profiles cannot be uniformly presented. Many continental and American workers have sought to describe profiles in terms of the total analyses of the individual horizons. Such figures, in the author's opinion, are not highly instructive since they include both the unweathered skeleton of the soil and the weathering complex, which is the seat of the chemical reactions of the soil. And thus an enrichment in sesquioxides in a given horizon, as shown by the total analyses, may either reflect the mechanical washing down of colloidal clay, or the translocation of sesquioxides from the clay complex of an eluvial horizon and their deposition in the horizon in question. The most desirable information would be in the form of analytical data for the composition of the clay complex in the different horizons of a profile. This information is given to some extent by certain workers who report constituents soluble in hydrochloric acid. If it could be assumed that the acid extraction used involved no attack on the unweathered minerals of the soil skeleton, such data would present a comparative view of the composition of the clay complex in the different horizons and would furnish information as to the character of the eluviation processes. Few soils, however, are free from original minerals susceptible to attack by hydrochloric acid. The error from this complication is likely to be greatest in relatively young soils. With certain tropical soils and soils whose parent materials have passed through long cycles of weathering, such as many of the Tertiary and Recent sediments, the proportion of unweathered silicates is likely to be less and the soil skeleton will consist mainly of quartz. Here the figures obtained by the use of hydrochloric acid extractions may have a real significance.

*Russia.** A sandy profile in the region of St. Petersburg was described by Georgievski³ as follows:—

*The profiles described in this and the ensuing chapters have been chosen to illustrate the distinctive features of the groups to which they belong. No attempt has been made to give examples from all countries.

Horizon A₁. Loose greyish-white layer, 4-5 inches thick.

Horizon A₂. White fine sand 10 inches or more thick.

Horizon B. A more or less compact or even hard mass of brown or black colour. In parts less cemented; thickness 5-10 inches.

Horizon C. Reddish-yellow or yellowish glacial sand.

In the same district, a clay soil gave :—

Horizon A₁. Whitish-grey fine grained layer, 5-6 inches thick.

Horizon A₂. Compact laminated almost white material, appearing white when dry and falling into a fine mealy powder. Some loam concretions present; thickness 3-5 inches.

Horizon B. Compact clayey material with abundant dark and brown concretions. The colour is mottled, whitish flecks alternating with reddish and yellowish streaks and threads of the slightly altered parent material; thickness about 8 inches.

Horizon C. Compact boulder clay of reddish-brown colour.

It will be noticed that in a heavier parent material the profile is markedly shallower. The tendency to laminated structure in the A₂ horizon is also to be noted and is a common characteristic of well developed podsoles.

It is to be presumed that, in the profiles just described, the B horizon is enriched both in organic matter and in sesquioxides. In certain sandy and gravelly profiles, the humus occurs as a distinct layer above the horizon of deposition of the sesquioxides.

Denmark. Podsol profiles under heath formation in Jutland, Denmark, have been exhaustively investigated by F. Weis⁴. Varying degrees of podsolisation are reflected in the changes in the composition of the inorganic colloids extracted by the ammonium oxalate method of O. Tamm⁵. Marked accumulation of humus with hardpan formation is shown in the B horizons of the most strongly podsolised profiles. Sesquioxide accumulation is most marked in the humus horizon, as shown by the composition of the ammonium oxalate extract.

England. W. M. Davies and G. Owen⁶ give the following description of a podsol profile at Goldstone, Newport, Shropshire, developed on Bunter pebble beds. This is the profile illustrated in the frontispiece :—

- A₀. 0-6 in. Leaf mould.
- A₁. 6-9 in. Greyish-black humified organic matter with bleached sand grains.
- A₂. 9-19 in. Light grey sand with bleached sand grains and an ashy appearance. Numerous pebbles free of iron oxide stains.
- B₁. 19-23 in. Humified organic matter compacted.
- B₂. Lower 1-4 inches of above. Layer cemented with iron oxide having a well defined ferruginous lustre.
- B₃. 23-35 in. Cinnamon yellow light sand with numerous stained pebbles. Decreasing intensity of coloration with depth. Somewhat indurated.
- C. Below 35 in. Pebbly brownish or purplish-red laminated friable sandy rock.

The principal analytical data are shown in Table XXVII.

TABLE XXVII.—ANALYTICAL DATA FOR GOLDSTONE PODSOL PROFILE (DAVIES AND OWEN).

Horizon	Clay %	Organic Carbon %	pH	Composition of clay fraction	
				SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
A ₁	2.7	14.88	3.70	—	—
A ₂	1.3	1.63	3.13	4.73	4.56
B ₁	8.8	7.14	3.73	1.56	1.48
B ₃	2.2	0.40	3.75	1.58	1.30
C	3.3	0.04	4.75	2.51	2.31

The above data give a clear picture of the nature of the process of podsolisation. In the first place, there is some evidence of mechanical eluviation in the rise in clay content in the B₁ horizon. Secondly, there is a marked rise in the organic carbon content in the B₁ horizon. Thirdly, the siliceous character of the clay fraction in the A horizons and the contrasted sesquioxidic character of the clay fraction in the B horizons should be remarked. Finally, the strongly acid reaction of the whole profile, and the increase in pH with depth is shown.

Scotland. Podsol profiles are of frequent occurrence in the Scottish Highlands, where they are developed under

heath or coniferous forest on igneous or metamorphic parent materials. A typical example on granitic drift in Aberdeenshire is described by W. G. Ogg.⁷ The rainfall is 30 to 40 inches per annum, and the vegetation coniferous forest.

- (1) Litter and blackish peat; thickness 5-7.5 cm.
- (2) Greyish bleached layer; thickness 5-7.5 cm.
- (3) Coffee-brown layer; thickness 1.25 cm.
- (4) Yellowish-brown non-cemented layer; thickness 20-25 cm.
- (5) Compact mottled and sometimes cemented layer.

It is to be presumed that (1) and (2) are A horizons, and (3) and (4) B horizons. The status of (5) is not clear, but it may possibly represent a transition to parent material with some illuvial accessions.

Wales. The following are typical of some Welsh podsol profiles, probably of secondary development from former B horizon material under heath vegetation.

Cefnybryn, W. Glamorgan. Altitude 300 feet. Mean annual rainfall, 40 inches. Mean annual temperature, 50°F. Parent material, Old Red Sandstone Conglomerate. Vegetation, *Ulex spp.*, *Festuca spp.* (described by D. O. Hughes and Brynmor Jones).

Horizon	Depth				Organic Carbon	pH
A ₀ .	0-4 in.	Fibrous peat layer	...	...	—	—
A ₁ .	4-7 in.	Greyish sand	...	...	1.61	4.6
B.	7-12 in.	Reddish-brown sandy loam	...	...	2.08	4.7
C.	14-16 in.	Yellowish-brown sandy loam on rock	...	...	0.48	5.5

The composition of the clay fraction in the several horizons is as follows:—

Horizon	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
A ₁ .	60.0	27.6	4.8	3.70	3.33
B.	43.0	30.0	21.7	2.44	1.67
C.	45.6	28.0	16.0	2.80	2.03

The data indicate a slight accumulation of organic matter and alumina and a marked accumulation of ferric oxide in the B horizon.

Aber, Caernarvonshire. Altitude, 900 feet. Mean annual rainfall, about 60 inches. Mean annual temperature, 47°F. Parent material, Ordovician shale. Vegetation, *Calluna vulgaris*, *Vaccinium myrtillus*, *Festuca spp.*, *Lichens*. (Described by G. H. Gethin Jones.)

Horizon	Depth		Organic Carbon	pH
A ₀ .	0-5 in.	Dark peaty layer ...	13.8	3.6
A ₁ .	5-8 in.	Grey stony loam ...	6.2	4.1
B ₁ .	8-14 in.	Reddish-brown loam ...	4.7	4.7
B ₂ .	14-20 in.	Light brown shaly loam ...	3.5	4.7
B ₃ .	20-26 in.	Light brown very shaly loam	3.6	4.8

The composition for the clay fraction in the several horizons is as follows:—

Horizon	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
A ₀ .	51.7	19.9	7.5	4.40	3.56
A ₁ .	45.4	20.1	13.1	3.84	2.75
B ₁ .	35.4	31.7	22.4	1.90	1.31
B ₂ .	34.7	28.3	22.3	2.09	1.39
B ₃ .	33.6	36.6	19.2	1.56	1.17

The data give no indication of accumulation of organic matter, but clearly show the deposition of ferric oxide and, to a smaller degree, alumina in the B horizons.

United States. Podsoles are of frequent occurrence throughout the more humid regions of the Northern United States and Canada. A profile collected by W. J. Geib⁸ near Bayfield, Wisconsin, is described as follows:—

- A₀. 0-3 in. Dark brown decayed leaves and other organic matter.
 A₁. 3-8 in. Grey with very slight pink tint.
 B. 12-30 in. Light to dark coffee brown, compactly cemented.
 C. 30-40 in. Pinkish-grey resting on heavy red clay.

The profile is developed in a fine sandy loam and shows a definite accumulation of clay in the B horizon. The silica-sesquioxide ratios of the colloidal clay in the successive horizons are 3.76, 4.01, 1.59, and 2.40 respectively.

W. J. Latimer⁹ describes the following profile from Washington, Massachusetts:—

- A₀. 0-6 in. Dark brown with mould and roots.
 B₁. 11-13 in. Dark brown or coffee coloured.
 B₂. 13-24 in. Yellowish-brown to pale yellowish-brown.
 C. 24-36 in. Yellowish-grey and streaked mottled brown.

The texture is a loam and increases in heaviness in the B and C horizons. The B₁ horizon shows an accumulation of organic matter. The silica-sesquioxide ratios of the colloidal clay in the successive horizons are 2.16, 2.74, 0.86, 1.28, and 1.67, respectively.

M. M. McCool, J. O. Veatch, and C. H. Spurway¹⁰, in studies on soil profiles in Michigan, have given the following generalised description of podsolised soils from the northern part of the State.

1. Thin humous soil (mould).
2. Grey podsolised soil (3-24 in.).
3. Brown horizon (dark coffee brown to light leather colour and dull yellow); thickness 4 in. to 4 feet; horizon of acid concentration.
4. Horizon showing iron oxide coloration; highest clay content; gradation to substratum.
5. Substratum.

Tropics. It was formerly considered that podsols were confined to temperate and cold humid regions. According to P. Vageler¹¹, however, podsolisation can occur to some extent in the tropics under conditions of extreme humidity. The natural vegetation is rain forest and the production of a podsol profile may intrude on an earlier stage of soil formation in which red loams have been formed. Unfortunately, no accurate descriptions of tropical podsols are at present available.

VARIATIONS IN PHYSICAL PROPERTIES IN PODSOL PROFILES.

The vertical variation in the physical properties of podsol profiles is governed by the texture of the parent material, the degree of development of the profile, and the thickness of the peaty layer. A. F. Lebedev and E. E. Bankova¹² have followed the vertical changes in a number of physical constants, including volume weight, porosity, and

maximum molecular water holding capacity for representative profiles. In general, the volume weight increases with the depth, but whilst in some cases it reaches a maximum in the B horizon, in other cases there is a continuous increase from the surface to the C horizon. The variations in porosity are generally in the opposite direction, a minimum occurring sometimes in the B, sometimes in the C horizon. Maximum molecular water capacity may increase with depth, and pass through a maximum in the B horizon or, in the case of humus podsols, show a maximum in the upper A horizon.

There is generally a certain amount of mechanical eluviation in podsol profiles. The accumulation of colloidal material in the B horizon is most marked in soils of light texture. This accumulation may be so pronounced in some cases as to lead to a certain amount of impedance of drainage with consequent waterlogging—a condition encountered in some of the podsol soils of the Bagshot area of Surrey. The development of an horizon with restricted permeability is doubtless a factor in the precipitation of sesquioxides and the development of a pan.

The degree of induration of the B horizon varies considerably. In some cases, although there is a definite horizon of accumulation, induration is only slightly developed. At the other extreme are rock-like B horizons such as occur in the Bagshot profiles. A pebbly or stony texture appears to favour the formation of highly indurated illuvial horizons.

In podsols developed from substrates of heavier texture, such as the shaly loam profiles of Wales, mechanical eluviation is less marked and the induration in the B horizon is comparatively slight.

MECHANISM OF PODSOLISATION.

We have already seen that podsolisation is a consequence of the removal of bases by leaching and the development of an acid organic layer at the surface. With regard

to the mechanism of the process, some difference of opinion exists. The accumulation of sesquioxides in the B horizon may reflect either a translocation and deposition of clay, which, relative to the coarser fraction of the soil, is rich in these constituents, or an actual differentiation of the clay complex due to a simple translocation of sesquioxides from the A to the B horizons. Whilst it is certain that in some cases mechanical eluviation, resulting in an actual transference of inorganic colloidal material from the A to the B horizon, takes place, the differences found in the composition of the clay complex in the several horizons by certain workers can only be explained by postulating a chemical eluviation.

The translocation and deposition of humus only occurs under very acid conditions. B. Aarnio¹³ has studied the mutual precipitation of humus and sesquioxide sols and finds that with ferric oxide-humus ratios varying from 1:0.2 to 1:2.8 precipitation occurs. The corresponding range for mutual precipitation of humus and alumina sols is 1:1 to 1:30. It can thus be seen that ferric oxide is more mobile in the soil profile than alumina, for when humus is in excess in the soil solution, at least 30 times as much humus as alumina must be present for this constituent to pass into colloidal solution. A more moderate concentration of humus suffices to ensure the translocation of colloidal ferric oxide. It is to be presumed that sols having the sesquioxides in excess of the humus do not occur under ordinary conditions. Assuming a sufficient solubility of the humus to give a humus-alumina ratio of over 30, then a decrease in this ratio owing to the removal of some of the humus will result in a mutual precipitation of the humus and alumina. Sufficient humus will remain in solution to protect the ferric oxide, which will be precipitated at lower levels.

The differentiation of the clay complex implied by the variations in its composition in the horizons of a podsol profile is a consequence of its instability under conditions of base-unsaturation. Where there is base-saturation, the clay

complex remains stable and no differentiation occurs. An essential for podsolisation is the presence of peaty organic matter such as occurs under coniferous forest or heath. It would appear that an acid reaction is not in itself sufficient to cause podsolisation. This may be readily shown in the laboratory by allowing 0.1 N-solutions of hydrochloric acid and oxalic acid, respectively, to percolate through ferruginous sand. Although the pH of the hydrochloric acid is lower than that of the oxalic acid, the solvent action of the latter acid is considerably greater and quickly leads to a bleaching by removal of ferric oxide. H. T. Jones and J. S. Wilcox¹⁴, from laboratory studies of the solution of sesquioxides in A horizons and their precipitation in B horizons, conclude that hydroxyacids play a part in the solution of sesquioxides, which are translocated in combination as complex anions and precipitated as basic salts.

The deposition of sesquioxides removed by leaching from the A horizon in podsol profiles has not yet been completely explained. There is generally an increase in base-saturation with depth, and this in itself would lead to some deposition of constituents which have been brought into solution by acid percolating waters. There is also the possibility that accumulation of sesquioxides may result, in part, from the restriction of free percolation in an horizon enriched in clay by mechanical eluviation.

C. G. T. Morison and D. B. Sothers¹⁵ consider the deposition of ferric oxide in iron pans to be due to irreversible precipitation of iron from humus-protected iron sols during periods of drought. During the winter months, the movements of soil moisture are mainly downwards and this results, if there is a surface layer of acid humus, in a downward leaching of ferric oxide, and, possibly, of alumina, in colloidal solution. With the drying out of the profile during summer, precipitation occurs. The localisation of this precipitation in a definite horizon is possibly connected with the course of the moisture gradient. It is known that capillary movement becomes negligible at moisture contents

in the vicinity of the moisture equivalent. Morison and Sothers consider the horizon of precipitation to be determined by the proximity of a level of complete saturation, but it does not appear possible to postulate a water table so near the surface. The horizon of deposition may occur at a depth where the subjacent strata are sufficiently moist to furnish a supply of precipitable material by capillary rise.

This view of the deposition of sesquioxides may apply also to the deposition of humic material. If it is correct, the B horizons are developed by precipitation not from descending solutions during periods of leaching but from ascending solutions during periods of drought.

BROWN EARTHS OR BROWN FOREST SOILS.

The name brown earths (Braunerde) was applied by Ramann to a group of soils occurring in Western Europe and developed naturally under deciduous forest. Their affinity with the podsoles consists in the complete leaching of carbonates from the mature soil profile and the development of a brown colour suggestive of the B horizon of the podsoles. They differ from the podsoles in some important respects, but chiefly in their better base status. This may be due to either or both of two causes. On the one hand, the climate may be such as to involve a less intense leaching, whilst, on the other hand, the parent material of the soil may be sufficiently rich in basic reserves to oppose the extreme impoverishment which favours podsol development.

The return to the soil of leaf-fall containing basic constituents is a powerful factor in maintaining the base status of these soils in their virgin state. The presence of a ground vegetation also assists in promoting the circulation of mineral nutrients through a deep layer of soil and subsoil. Unlike the soil under coniferous forest, the soil under deciduous forest maintains a population of earth worms, which are the most important agents in distributing organic matter throughout the layers which they occupy. And thus, whilst the organic matter profile under conifers consists of a

sharply defined peaty layer, overlying a mineral soil relatively poor in organic matter (although sometimes including a humus B horizon), the organic matter of the soil of a deciduous forest shows a fairly uniform distribution throughout the top foot or eighteen inches, decreasing gradually in deeper layers. An accumulation of organic matter in a B horizon such as occurs in the humus podsols is never found in the brown earths.

The marked translocation of sesquioxides, which is betokened by the bleaching of the A horizon and the brown or reddish-brown coloration of the B horizon in podsols, is considerably less marked in the brown earths, where bleaching is scarcely shown. Apart from mechanical translocation, the clay complex remains comparatively unaltered throughout the profile.

A. Stebutt¹⁶ considers that, in the brown earths, the clay complex has undergone a certain degree of degradation into its constituent silicic acid and sesquioxides. The brown colour of these soils is an evidence of the presence of free hydrated ferric oxide. A further stage in the degradation process consequent on impoverishment in bases would involve the differential leaching out of the sesquioxides and the development of a bleached A horizon as in the podsols.

Structurally, the brown earths show important differences from the podsols. Whilst in the podsols the single grain structure predominates, the brown earths are generally distinguished by a crumb structure or a variant of this structure. This may be attributed to the better base status and to the better distribution of base-saturated organic matter, throughout the soil layers.

Sweden. The character of the brown earth profile is illustrated by the following Swedish profile described by K. Lundblad¹⁷. It was found under beechwood.

- A₀. Beech litter, 2-3 cm. thick.
- A₁. Loamy layer containing mild humus with indefinite lower limit. Thickness 12-15 cm.
- B. Brown earth. Thickness 40 cm.
- C. Sandy, stony, moraine, principally of granite material.

Wales. The characters of a Welsh brown earth profile are shown in the following description :—

Llanfairpwll, Anglesey. Altitude, 120 feet. Mean annual rainfall, 38 inches. Mean annual temperature, 50°F. Parent material, Pre-cambrian schistose drift. Artificial grassland, but originally deciduous forest.

Horizon	Depth	Description	Organic Carbon	pH
I.	0-9 in.	Dark brown stony loam ...	2.9	5.40
II.	9-15 in.	Brown stony loam ...	2.05	5.64
III.	15-36 in.	Reddish-brown stony loam...	1.1	5.94
IV.	36-48 in.	Yellowish-brown stony loam	nil	5.80

There is no well marked differentiation between A, B, and C horizons. The variations in the composition of the clay fraction are shown in the following table :—

Horizon	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
I.	43.9	31.2	16.1	2.39	1.80
II.	45.1	31.4	16.7	2.47	1.85
III.	44.8	31.6	16.6	2.41	1.80
IV.	45.6	29.4	13.2	2.64	2.05

These figures exhibit a reasonable constancy in composition of the clay fraction in the first three layers. The fourth layer may well represent C horizon material. So far as an A horizon is concerned, either it has been removed by erosion or else it has been mixed by cultivation with the B horizon. Neither the field character nor the laboratory data offer any evidence of an eluviated horizon.

England. The above profile is probably typical, in its essential features, of large areas of similar soils in England and Wales, and may be expected in well drained situations with parent rocks of moderate base status under British climatic conditions. Where, as in certain coarse sands, notably in the Trias, the base reserves are poor, podsoles are developed. It is thus possible for brown earths and podsoles to occur under exactly similar climatic conditions, the difference in the final result being due to the nature of the parent material.

N.W. Europe and N. America. Brown earths occur commonly over a large area of Western Europe. They occur also in the Eastern United States from New York to N. Carolina. The American brown earths often show a marked development of a texture profile due to the mechanical eluviation of clay and fine material from the surface horizons. From the scanty data available it would appear that there is no marked difference in the composition of the clay fraction in the different horizons, as would be the case if chemical eluviation had been operative. The removal of clay from the surface horizons does, however, produce a weakening of the soil colour in the top soil, so that many of these soils appear somewhat bleached. It should be added that in many districts there has been extensive sheet erosion, with the result that the original B horizon is now exposed, and in many cases forms the cultivated soil.

ENGLISH CULTIVATED SOILS AND FOREST SOILS.

Most of the soils of England with good natural drainage were probably brown earths under primitive conditions, but owing to their long agricultural history, they must be regarded as highly artificial. The original deciduous forest which covered most of the country has been almost entirely removed and authentic natural profiles can only be studied in isolated localities such as the Forest of Dean, the New Forest, and the Forest of Wyre. It is to be regretted that, at present, no data are available for profiles in these areas.

Under ordinary farming conditions, the incorporation of organic matter with the soil by the addition of farmyard manure and by the cultivation of artificial grass leys, and the maintenance of the base status by chalking, liming, or marling, have tended to modify the original brown earth soil and produce a type of soil which resembles the black earths in its content of well distributed organic matter, its high base status, and the absence of sesquioxide eluviation.

We have seen that brown earths are developed under deciduous forest and that the relatively high mineral content //

of the leaf litter and the presence of a gramineous and herbaceous ground vegetation maintain the base status of the soil and oppose podsolisation. Where the natural deciduous forest is removed and the soil is afforested with conifers, it may happen that deterioration in base status proceeds to such a degree as to result in incipient podsolisation. This may be seen in plantations on light soils in England, for example, in the Delamere Forest of Cheshire. Similar deterioration takes place where natural deciduous forest becomes replaced by heath. Podsolisation is a deteriorative change, and it is important to realise that, on light soil under heavy rainfalls, the change from brown earth to podsol may easily follow the removal of deciduous forest.

PRAIRIE SOILS.

The podsoles and their congeners are for the most part developed under forest. Beyond the forest limit to the north lie the tundras, where, although the humidity is moderate or excessive, the general rigour of the climate precludes forest growth. There is an important group of soils, lying on the arid side of the eastern forest belt of the United States and occupying a considerable proportion of the so-called Middle-West. These soils, although developed under steppe or prairie conditions, appear to belong to the great group of humid soils rather than to the arid soils, which lie adjacent to them and are also developed under steppe conditions. The rainfall under which the prairie soils occur is sufficient to produce a complete leaching of soluble salts, including carbonates, in the soil profile, and there is thus no horizon of accumulation of calcium carbonate or gypsum as in the neighbouring black earths. Calcium carbonate encountered in a prairie profile belongs to the C horizon and is to be reckoned with the parent material. A characteristic prairie soil profile consists of three horizons, namely, a dark coloured horizon rich in organic matter, a transition horizon often showing brown colours, and the underlying parent material, which is generally of a grey

colour. Calcium carbonate is found at a depth of 50-60 cm. where, as is frequently the case, the parent material is a calcareous loess.

The following profile at Fargo, N. Dakota, is typical :—
A(B). 0-50 cm. Dark granular humous horizon sharply differentiated from
C. Porous loess with carbonates and rusty flecks.

The granular structure is not so marked as in the black earths and may be entirely lacking, whilst the calcium carbonate in the C horizon is quite distinct from the concretionary deposits found in the more arid soils.

The prairie soils resemble the black earths in their organic matter profile. The abundant and deep rooted grass vegetation, the presence of a rich soil fauna, and, in spite of the leaching of calcium carbonate, the favourable base status, all result in a deep enrichment of the soil in dark coloured base-saturated organic matter.

Eluviation of sesquioxides and deposition in a B horizon does not occur. It may be conjectured that the clay complex in prairie soils has scarcely suffered to any appreciable extent that degradation into free silicic acid and sesquioxides which Stebutt postulates for the brown earths. The dark colours of the prairie soils mask field evidence of changes affecting the inorganic complex, but incipient degradation is sometimes shown by the presence of brownish colorations below the dark coloured horizon.

It should be added that H. Stremme¹⁸ and certain other continental writers consider the North American prairie soils as " meadow " soils whose distinctive characters are due to the presence of ground-water. On the other hand, N. Florov¹⁹ assigns these soils to the class of degraded tsher-nosems (see page 219). The American workers class them definitely with the humid soils and relate them to the podsol group. It is considered that the prevalence of steppe rather than forest conditions is due to the prevention of natural regeneration through grazing animals and prairie fires. It is certain that, since the extinction of the herds of

bison, and the prevention of fires by settlement, forest vegetation has tended to invade these regions, and it may be reasonably conjectured that, in the absence of interference, the prairie soils would undergo a kind of degradation analogous to that exhibited in the forest soils bordering on the tshernosem regions of Russia. (See also p. 245.)

Prairie soils doubtless occur in many parts of the world on the humid side of the boundary between the humid and arid soils. Economically, they must rank among the most fertile of the great groups, and their recognition and definition is an important task for the future.

There is a class of soils occurring in the humid west of Britain which may be considered to have some affinities with the prairie soils. These are the soils of permanent grasslands. The resemblance consists in their high content of organic matter well distributed throughout the soil profile. But, inasmuch as, except on calcareous soils, their base status is low, the type of organic matter formed by microbiological decompositions will be greatly different, being of a less marked black colour. Further, the clay complex under such conditions is less stable than in the prairie soils and may undergo degradation, as in the brown earths, or even differentiation if the profile is sufficiently acid to permit podsolisation.

DEGRADED TSHERNOSEMS.

We have seen that the tshernosem belt of N. America is succeeded on the side of increasing humidity by the so-called prairie soils. The European tshernosem passes into a belt of soils which are described as degraded tshernosems or grey forest soils. It is considered that the region occupied by these soils represents former steppe and that the profiles found there have certain surviving characteristics of the black earths.

In the first stages of degradation under forest, the changes are mainly structural and consist in the destruction of the characteristic porous crumb structure of the

tshernosem. There is also a sinking of the calcium carbonate horizon and some loss of organic matter. In a more advanced stage of degradation, there is a further obliteration of the porous structure and the development of a mineral profile, characterised by eluviation and deposition of sesquioxides and the disappearance of carbonates from the profile, resulting in the formation of the so-called grey forest soil.

The following data by Georgievski are quoted by Glinka²⁰ for a grey forest profile from the Poltava government:—

- A₀. 0-2.5 or 5 cm. Forest litter.
- A₁. Dark brown to light grey granular layer, the granules increasing to walnut size, down to 26 cm., where there is a change to
- A₂. Ash grey horizon with nut structure. On drying, breaks up into small angular blocks with greyish-white powdery coating. The thickness of this horizon is 47-48 cm.
- B₁. Reddish-brown compact loam at first containing organic matter and showing nut structure. Dark brown deposit in pores and cracks. The thickness of this horizon is 70-140 cm.
- B₂. Brownish, highly calcareous loam sometimes passing into hard white marl. Thickness 70-140 cm.
- C. Yellow loess.

A transition occurs to typical podsol soils on the side of increasing humidity.

The decisive feature of the degraded tshernosems or grey forest soils is the development of a brown or reddish brown colour in the subsoil horizons, indicating the operation of podsollic eluviation. The decrease in organic matter content compared with adjacent black earths has been shown by numerous analyses. It appears, however, that the organic matter profile shows a more gradual decrease of this constituent with increasing depth than in the typical podsols.

The Russian workers have recognised a number of stages in the degradation of tshernosems, ranging from the unchanged tshernosem to definite podsollic forest soils with markedly bleached horizons below the forest litter. H. Stremme²¹ and others have described soils in Germany which

appear to belong to the class of degraded tshernosems. Examples are given by C. E. Kellogg²² from Wisconsin, where grey forest soils appear to have developed from former prairie soils.

MOUNTAIN SOILS.

The soils of alpine regions resemble soils of high latitudes, principally in the circumstance that, owing to the prevalence of low temperatures, weathering is mainly of a physical character. Apart from the low temperatures under which they are developed, the most important factor affecting them is the strong relief, which facilitates erosion and, except in valleys, prevents the accumulation of deep layers of weathered material.

Generally speaking, there is a high precipitation in mountain regions. This is not invariably the case, and it may also happen that the rainfall does not increase uniformly with altitude, but passes through a maximum at intermediate altitudes. The humid character of mountain regions favours intense erosion, and the soils on hillsides are generally thin, whilst valleys, if not occupied by glaciers, are deeply filled with the spoil of the higher lands. The intense leaching results in removal of bases so that soils are generally acid in character. H. Jenny²³ found that, even on calcareous rocks, about 80% of alpine soils examined in Switzerland were actually acid. The acid character of the soil and the low temperatures which prevail favour the accumulation of peaty organic matter, the so-called alpine humus, in folds and hollows. Generally, however, erosion prevents the accumulation of more than about 4 inches of rocky humus soil.

The relief is generally the most important factor in differentiating soils of mountain regions. Thus, whilst in exposed situations the soil may be represented by bare rock or spreads of angular rock fragments, on gentle slopes and shelves there may be a sufficient depth of soil to carry a closed association of grass—the alpine meadow, or at lower elevations, forest.

Alpine humus soils occur above the level of forest. Whilst humidity generally increases with altitude, at great elevations, maximum humidity is passed and a decrease sets in. This decrease may in some cases determine the recession of forest growth and the prevalence of grass vegetation. In other cases, the limit may be imposed by the rigour of the climatic conditions. The height at which forest gives place to alpine humus varies with latitude. In Britain the limit may lie at 2000-2100 feet, whilst in equatorial regions it may be as high as 13000-14000 feet.

The profile is relatively simple and consists of a layer of humus soil overlaying rock. The thickness is variable but generally small. The organic matter content varies considerably. Whilst in some cases the soil may consist entirely of a felt of fibrous peaty matter, in other cases only a few per cent. of organic matter is present.

The vegetation of alpine humus soils is of a decidedly xerophilous character. The conditions, however, differ markedly from those obtaining in deserts, for the adaption is not to predominantly arid conditions but to strongly variable conditions, which include periods of intense drought. The drought is intensified by the thinness of the soil, the prevalence of high winds, and the strong insolation. Xerophytic characters can be observed also in heath vegetation at lower altitudes.

The soils in wet hollows at high altitudes are naturally of a peaty character. Thick layers of peat may be developed where slopes are maintained constantly wet by surface water.

Mountain soils are generally dark in colour owing to the presence of organic matter. The colour of the mineral soil is generally grey. Owing to the low temperatures, there is no chemical weathering and thus no formation of colloidal clay to undergo degradation as in the podsols. Whilst clay of primary origin is not likely to occur in high alpine regions, it may happen that, where the parent rock consists of unmetamorphosed sediments, soils containing colloidal clay will occur.

Owing to their inaccessibility, comparatively little is known as to the conditions of soil formation at high altitudes. The altitude at which the distinctive characters of mountain soils occur varies, of course, with latitude, and whilst in Central Africa these characters are only seen near the peaks of the highest mountains, in high latitudes they are encountered at comparatively low elevations. Indeed, the principal distinction between alpine soils and tundra consists in the nature of the relief forms.

BIBLIOGRAPHY.

- ¹ SSUKATSCHEV; from K. D. GLINKA, *Die Typen der Bodenbildung*, Berlin, 1914, p. 173.
- ² NIKIFOROV, C., "The perpetually frozen subsoil of Siberia," *Soil Sci.*, 1928, **26**, pp. 61-77.
- ³ GEORGIEVSKI; from K. D. GLINKA, *Die Typen der Bodenbildung*, Berlin, 1914, p. 68.
- ⁴ WEIS, F., "Physical and chemical investigations on Danish heath soils." *Kgl. Danske Vidensk. Selsk. Biol. Medd.*, 1929, **7**, pp. 1-196.
- ⁵ TAMM, O., "Markstein i det nordsvenska barr skogs-området." *Medd. fr. Stat. Skogsfors.*, 1920, **17**, p. 49.
- ⁶ DAVIES, W. M., and OWEN, G., Private communication.
- ⁷ OGG, W. G., "Scottish soils in relation to climate and vegetation." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. I, pp. 301-309.
- ⁸ GEIB, W. J.; cited by ANDERSON, M. S., and BYERS, H. G. "Character of the colloidal materials in the profiles of certain major soil groups." *U.S. Dept. Agr.*, 1931, Techn. Bull. 22.S.
- ⁹ LATIMER, W. J.; cited by ANDERSON, M. S., and BYERS, H. G., *ut supra*.
- ¹⁰ MCCOOL, M. M., VEATCH, J. O., and SPURWAY, C. H., "Soil profile studies in Michigan." *Soil Sci.*, 1923, **16**, pp. 95-106.
- ¹¹ VAGELER, P., *Grundriss der tropischen und subtropischen Bodenkunde*, Berlin, 1930.
- ¹² LEBEDEV, A. F., and BANKOVA, E. E., *Physical characteristics of the soil profile*, Moscow, 1930.
- ¹³ AARNIO, B., "Experimentelle Untersuchungen zur Frage der Ausfällung des Eisens in Podsolböden." *Int. Mitt. Bodenk.*, 1913, **3**, pp. 131-140.
- ¹⁴ JONES, H. T., and WILCOX, J. S., "Studies in soil genetics." *I. J. Soc. Chem. Ind.*, 1929, **48**, pp. 304-308.
- ¹⁵ MORISON, C. G. T., and SOTHERS, D. B., "The solution and precipitation of iron in the formation of iron pan." *J. Agric. Sci.*, 1914, **6**, pp. 84-96.
- ¹⁶ STEBUTT, A., "Die Braunerde (ein Beitrag zur Theorie der Braunerdebildung)." *Z. Pflanz. Düng.*, 1930, **15A**, pp. 134-167.

- 17 LUNDBLAD, K., "Bidrag till kannedomen om brunjordseller mulljords-
typen egenskaper och degeneration i sodra Sverige." *Medd. fr.
Statens Skogsförsöks-anstalt*, 1924, **21**, pp. 1-48.
- 18 STREMMER, H., "Die Prarieböden," in E. BLANCK'S *Handbuch der
Bodenlehre*, Band III., pp. 287-294.
- 19 FLOROV, N., "Zur Frage der Degradierung der dunkelfarbige Böden
von Nord-Amerika." *Soil Res.*, 1929, **1**, pp. 200-224.
- 20 GEORGIEVSKI; from K. D. GLINKA'S *Die Typen der Bodenbildung*,
Berlin, 1914, p. 90.
- 21 STREMMER, H., *Grundzüge der praktischen Bodenkunde*, Berlin, 1926,
pp. 102-110.
- 22 KELLOGG, C. E., "Preliminary study of the profiles of the principal
soil types of Wisconsin." *Wis. Geol. Nat. Hist. Survey*, 1930,
Bull. 77A, pp. 1-112.
- 23 JENNY, H., "Die alpinen Böden." *Denkschr. schweiz. naturforsch.,
Gesell.*, Zurich, 1926.

CHAPTER X.

TSHERNOSEMS AND THEIR RELATED GROUPS.

THE soils to be described in the present chapter are developed in semi-arid and arid climates with incomplete leaching which results in the development of horizons of carbonate deposition. They are developed under steppe or desert vegetation. The term *pedocal*, suggested by Marbut, may be conveniently used to cover the groups comprised in the succession from black earths to grey-brown desert soils.

TSHERNOSEMS* OR BLACK EARTHS.

The study of the tshernosems or black earths is of considerable interest and importance, not only on account of their wide extent and agricultural value, but because the modern study of soils in their natural relationships received its greatest impetus in the investigations and discussions of the problems of the origin of these and related soils. Further, regarding the diversity of the soils of the world as resulting from the varying intensity with which the pedogenic processes depending on temperature and moisture have operated, it is apparent that the black earths occupy, as it were, a central position between extremes with respect to the intensity of the factors which have affected their development.

Russia. The obvious feature of the black earths is, as the Russian name implies, their colour, which is due to the presence of humified organic matter. In the centre of the

*The transliteration of Russian names and words always presents difficulties. In a series of papers by K. K. Gedroiz, translated by Dr. S. A. Waksman for the U.S. Bureau of Plant Industry, the present form is given, on the authority of Dr. Gedroiz, as preferable to such forms as chernosem, tshernoziem, and tshernosiom.

tshernosem area in Russia, where the conditions are most typical for the development of these soils, the colour is deep black, but as the adjoining soil provinces are approached, the black colour becomes less marked, changing into the grey of the degraded tshernosems on the one side and the brown of the chestnut earths on the other.

The intensity of the dark colour produced would lead an observer familiar only with soils of humid climates to expect larger proportions of organic matter in the tshernosems than actually occur. A sample of tshernosem from the Ukraine, examined by the writer, contained only 3.2% of organic carbon, i.e., about 6% organic matter; yet the dark colour suggested a much higher proportion. As much as 16% of organic matter may be present in the central tshernosems, but 8 to 10% is more usual. The dark colour may be attributed to some extent to the fact that these soils are base-saturated, and that the humification processes proceed in a neutral or even alkaline medium. It would seem, however, that other factors must be involved. Certain Lower Lias soils in Britain contain free calcium carbonate and notable proportions of organic matter, yet do not exhibit the dark colour of the tshernosems. In all probability the high temperature to which these soils are subjected during summer, in conjunction with the high base status, may determine their dark colour. It is noteworthy that, in many parts of the tropics and sub-tropics, dark coloured soils are associated with a high base status, whilst base-unsaturated soils, under similar climatic conditions, show little or no humic colour, although they may actually contain higher proportions of organic matter than neighbouring dark coloured soils.

The thickness of the dark coloured layer in the tshernosems varies considerably. The average for the best developed black earths is about 70 to 100 cm., with 40 cm. to 150 cm. as extremes. It is considered to be differentiated into two horizons, an upper, A_1 , and a lower, A_2 , the latter having a lighter colour and a lower organic content. The

thickness of the dark horizons and the organic matter content decrease towards the boundaries of the soil group. The sharpness of the transition to the underlying light coloured parent material varies somewhat, but the change generally takes place within a few centimetres.

With regard to the inorganic profile, the typical tshernosem shows complete leaching of soluble salts, incomplete leaching of calcium carbonate and calcium sulphate, and no leaching of silica or sesquioxides. Thus, whilst there is no accumulation of sodium salts in the profile, an unfailing mark of the tshernosem is the zone of accumulation of calcium carbonate, sometimes accompanied by calcium sulphate.

The depth of this zone will vary according to the parent material. It approaches the surface more nearly the more calcareous is the parent material and also the more arid is the climate. The calcium carbonate zone may occur in the humic layer, but is generally near or below its lower limit. Above the main calcium carbonate layer, calcium carbonate occurs deposited in fibrous threads, like fungal hyphæ, and in small concretions which are sometimes hollow and may assume curious forms. Gypseous deposits may occur below the calcium carbonate horizon.

In some cases, a second horizon of calcium carbonate accumulation may be observed at a depth of a few metres, but this would not appear to be a necessary characteristic.

A characteristic peculiarity of tshernosems, and indeed of the adjacent chestnut coloured earths, is the presence of *krotowinas*. These are formed by the filling in by calcium carbonate of the burrows of steppe animals.

The black earths are characterised by their well developed granular crumb structure in the humic layer. This is, of course, associated with their high base status and the tshernosem is thus the chief example of a "calcium soil." Although these soils may contain high proportions of clay, they offer a few difficulties in tillage and consistently maintain their crumb structure.

So far as the clay complex of the soil is concerned, there would appear to be no vertical differentiation. It is unfortunate that no analyses of the clay fraction of Russian black earth profiles are available; but there is no evidence from field observations or from laboratory data to suggest the eluviation of sesquioxides or silicic acid. Neither is there any evidence of the eluviation of the clay complex as such, with consequent development of a texture profile. It is held by A. Stebutt¹ that, in the tshernosems, the conditions for the formation and maintenance of the aluminosilicic weathering complex are at optimum. Increase of temperature and increase in humidity both favour the destructive processes whereby the component silicic acid and sesquioxides are liberated and, in the podsoles, differentiated.

The brown or reddish-brown colours associated with the presence of free sesquioxides are entirely lacking in the typical tshernosem profile, in which the mineral colour is predominantly greyish-brown. The transition to degraded tshernosems in the direction of increased humidity is marked by the appearance of brown sub-soil horizons. The transition to higher temperature conditions is marked also by the development of brown colours throughout the profile, indicating the destruction of the weathering complex without eluviation.

In Russia, the black earth region is characterised by cold winters and hot summers, with a moderate rainfall. Data presented by P. Kossowitsch² show mean January temperatures from -20.8°C in Siberia to -3.5°C in the Ukraine, and mean July temperature varying from 16.7°C in Siberia to 23.3°C in S. Russia. The mean annual rainfall varies from 300 mm. to over 500 mm. A considerable proportion of the precipitation is in the form of snow, and the summer rainfall is heavily outweighed by evaporation. These conditions result in steppe vegetation which grows luxuriantly in spring and in early summer. The ensuing drought puts an end not only to active plant growth but also to microbiological activity in the soil.

In Russia and Asia, where the black earths have been principally studied, the parent material is generally loess and the textural range is limited by the character of this deposit. The extent of the loess and of the tshernosems are, however, not necessarily identical, for not only do other soil groups occur on the loess, but tshernosems may also be developed from other parent materials.

K. D. Glinka³ describes the following profile from Kogalnitza, Rostov :—

- A₁. 0-20 cm. Black with brownish-grey shading, breaking down into irregular lumps and eventually into a porous fine grained material. Effervescence with HCl from 8 cm.
- A₁. 20-55 cm. Greyish-black granular; more cohesive below. Columnar and nut structure.
- A₂. 55-90 cm. Gradual transition to greyish-brown mottled appearance. Nut structure. Krotowinas. Structural elements with white coating of carbonate.
- A₂. 90-105 cm. Disappearance of dark colour and irregular transition to brownish-yellow clay. Krotowinas. Calcium carbonate concretions.
- A₂-C. 105-185 cm. Yellowish-brown clay with vertical tongues of humous material. Calcium carbonate concretions.
- A₂-C. 185-240 cm. Humous streaks in wet pores. Decrease in carbonate concretions.
- C. Yellowish-brown clay.

Germany. Soils which appear to belong to the black earths occur in many parts of Germany. The following profile is described by K. Von See⁴, from near Magdeburg :—

- A. 0-55 cm. Humous carbonate-free loess. Loose and porous. Below the root horizon down to 30 cm. a greyish-brown horizon with lump structure over an almost black humous layer with tendency to polyhedral structure. Occasional flecks of greyish-yellow carbonate-free loess. At the transition to parent loess a zone rich in carbonate.
- C. Yellow or yellowish-grey unstratified loess.

V. Hohenstein⁵ describes numerous tshernosem profiles developed on boulder marls, clay marls, and glacial sands, in Eastern Germany. They are characterised by the leaching of carbonate from the upper horizons down to 20-100 cm. and the development of a zone of carbonate accumulation. The humus content, 2-4%, is rather less than in the Russian

black earth region. The prevailing colours are coffee brown to dark brown. They are considered to have been formed under an earlier steppe climate.

Czechoslovakia. V. Novák⁶ describes "Hanna" soils in Czechoslovakia which appear to belong to the tshernosem group. A typical profile from Hulin showed:—

- 0-40 cm. Dark loam with brownish tinge. Loose crumb structure sharply delimited below.
- 40-70 cm. Compact humous loam easily breaking down into crumbs.
- 70-100 cm. Dark compact loam.
- 100-150 cm. Yellowish loam.
- 150-350 cm. Light coloured loam with concretions and superficial deposits of calcium carbonate.

The mechanical texture is very uniform. The humus content decreases gradually from 4.83% in the surface horizon to 3.67% at 100 cm. There is then a sharp fall to 2.00% at 120-140 cm.

Another type of Hanna soil differs from that described above in being lighter in colour and poorer in humus. Calcium carbonate concretions appear at 130-140 cm.

North America. Tshernosems occur over considerable areas of North America in the States or Provinces of Saskatchewan, Alberta, Minnesota, the Dakotas, Nebraska, Kansas, Colorado, Oklahoma and Texas. Whilst the northern soils resemble the Russian tshernosem, some change in character occurs in passing to the warmer conditions of lower latitudes.

A northern tshernosem from N. Dakota, mapped as the Barnes series, is thus described by C. F. Marbut⁷:—

- I. 0-12 in. Heavy black loam. Irregular clod structure in top 4 in., ill-defined columnar structure below.
- II. 12-22 in. Dark greyish-brown clay loam with ill-defined columnar structure. No effervescence with acid.
- III. 22-32 in. Light greyish-yellow to greyish-brown, heavy silt loam or silty clay loam. White spots and splotches of calcium carbonate. Columnar structure disappears in this horizon, which is marked by maximum of carbonate accumulation.

- IV. 32-40 in. Mottled silty clay loam. Small iron oxide stains and concretions present. Spots and concretions of calcium carbonate less abundant than in III.
- V. Greyish-brown silty clay loam with iron oxide stains and concretions. Calcium carbonate present in disseminated form. This is the parent glacial drift from which the profile is developed.

The following is Marbut's description of a southern tshernosem from Texas, mapped as the Amarillo series:—

- I. Yellowish laminated fine sandy to silty loam becoming darker coloured downwards. Presence of æolian material in surface. Grass roots abundant below. Thickness about 4 in.
- II. Dark brown to very dark brown, generally clay loam. Dark yellowish-brown or reddish-yellow on being crushed. Granular structure becoming coarser with depth. Thickness about 12 in.
- III. Reddish-yellow clay loam, still with darker coloured coatings in structural elements. General tendency to columnar structure.
- V. Reddish-yellow material with much calcium carbonate. Similar structure to III.
- ✓. Parent material; unconsolidated calcareous silt and clay.

Data from M. S. Anderson and H. G. Byers⁸ for a Texas profile, which, though somewhat different from the Russian tshernosem, is essentially similar in its general type of development, are given in Table XXVI. The soil series is the same as that described above.

TABLE XXVI.—COMPOSITION OF COLLOIDAL MATERIAL OF A PROFILE FROM THE AMARILLO SERIES.

Depth	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /R ₂ O ₃
0-5 ins. ...	50.51	22.04	8.80	3.10
10-20 ins. ...	51.51	22.71	8.61	3.09
30-40 ins. ...	51.32	22.43	8.46	3.13
54-64 ins. ...	51.23	24.08	8.19	2.97
70-75 ins. ...	38.42	17.64	5.71	3.06 CaCO ₃ horizon
96-100 ins.	45.88	20.09	6.89	3.18

The composition of the clay colloid shows a remarkable constancy throughout the profile and there is no evidence for eluviation either of silicic acid or sesquioxides.

H. W. Hawker⁹ describes a range of soils in the Rio Grande valley of Texas which show successive stages of leaching from young delta soils to mature terrace soils with calcium carbonate accumulations at 4 to 8 feet. The mean annual rainfall is 22.5 inches and the profile consists of light brownish-grey to greyish-brown soil overlying heavy bluish-grey clay, mottled yellow and, sometimes, red. The surface horizons are non-calcareous.

India. There is in India a well known type of soil known as the *regur*, or *black cotton soil*, which occupies a large area in the southern half of the Peninsula. The natural vegetation appears to be of a grass steppe type. Regur is usually of a clay character and varies in depth from 1 to 2 metres, occasionally reaching 5 metres. On drying it shrinks and cracks to such an extent that it is often said that "black soil ploughs itself." Like the tshernosem, it is characterised by a zone of calcium carbonate concretions ("kunkur"). The parent material is generally a black basalt trap, the so-called Deccan trap. The organic matter content is generally low and the black colour, according to H. E. Annett¹⁰, is due in part to the presence of finely divided titaniferous magnetite.*

The climatic conditions for the Indian regur soils differ from those in the Russian tshernosem region in that the temperature, even in the winter months, is always considerably higher. The rainfall, rather more abundant than in the Russian black earth region, is, however, insufficient to compensate for the great evaporation resulting from the high temperatures which prevail, and incomplete leaching results. The climate is thus not identical with that of the Russian tshernosem region. Regur might, therefore, be described

*W. H. Harrison and M. R. R. Sivan (Mem. Dept. Agr. Ind. Chem. Series, 1912. II. No. 5) only admit Annett's explanation for the trap areas. Elsewhere, colloidal iron and aluminium silicates are held responsible for the black colour.

as tropical black earth. Black soils occur in Morocco, Kenya, and other parts of Africa, which may have affinities with the black earths, and are, probably from their resemblance to regur, also termed "cotton soils."

Argentina. The soils of the Argentine Pampas would appear also to present many similarities with the black earths and further work may result in their being assigned to this group.

The black earth represents the soil type developed under the conditions which prevail on the arid side of the boundary between humid and arid soils. When they have been completely investigated in all parts of the world where they occur, it will probably be possible, as Marbut suggests, to distinguish them into sub-groups on the basis of the prevailing temperatures. We may thus speak of high latitude, mid-latitude, sub-tropical, and tropical tshernosems. The common characteristics are a dark A horizon, relatively rich in base-saturated organic matter, a high base status, and the presence of an horizon of calcium carbonate accumulation. The vegetation is a closed association of steppe, savanna, or pampas character, either devoid of trees or else with isolated individuals as in the S. African bush veldt.

BLACK COTTON SOILS OF KENYA.

Grey and black soils occur in Kenya which may represent tropical tshernosems. D. S. Gracie¹¹ describes a group of soils, locally termed "black cotton soils," which appear to exhibit various stages of degradation. In the undegraded type, the profile consists of a black clay, rich in exchangeable calcium, overlying a layer of calcium carbonate accumulation. In the degraded soil, there is a grey layer at the surface with acid reaction and low exchangeable calcium content. This overlies the black clay. A still further stage of degradation is seen in certain soils with acid greyish-brown surface horizons. Degradation appears to involve a loss of organic matter. Black cotton soils may also be developed on lighter textured parent materials. They may be

analogous with the vleis soils (see p. 243) of the humid regions of Kenya, and tend to occur in plains and depressions in relatively hot and dry regions. Their complete analogy with the tshernosems of higher latitudes would appear to be still unproven.

H. H. Bennett and R. V. Allison¹² report soils with zones of calcium carbonate accumulation in the drier parts of Cuba, but doubt their analogy with the tshernosems of the United States.

CHESTNUT-COLOURED SOILS.

The group known as the chestnut-coloured soils (also called chestnut earths) occurs in Russia and Asia on the arid side of the region of the related tshernosems. They differ from tshernosems principally in the lower proportions of organic matter in the humous layer and in the closer approach to the surface of the zone of carbonate accumulation. Both circumstances are intelligible consequences of the drier character of the climate as compared with that of the tshernosem zone.

According to K. D. Glinka¹³, the A₁ horizon consists of an upper portion (5-7 cm.) of light colour and loose stratified structure, and a lower portion of smaller thickness and compact structure, lacking the characteristic granular character of the tshernosems. The A₂ horizon, which may reach 60 cm. in thickness, is lighter in colour and compact. Both horizons show a prismatic structure by reason of the occurrence of vertical cracks. The organic matter content of the humous horizons varies from 3 to 5%. The mechanical composition is fairly uniform and there is probably no vertical differentiation of the constituents of the clay complex. Accumulation of calcium carbonate may occur in the A₁ horizon or even near the surface. Gypsum may occur in the lower part of the A₂ horizon.

The colours of the humous horizons are greyish-brown and do not appear to be influenced by the presence of free hydrated ferric oxide.

The natural vegetation of the chestnut earths is low-grass steppe.

BROWN AND GREY SOILS OF THE SEMI-DESERT.

With passage to more arid conditions, the chestnut-coloured soils are replaced successively by brown and grey desert soils. These differ from the chestnut coloured soils by their lower content of organic matter and by the closer approach to the surface of the calcareous and gypseous horizons.

United States. In a study of soil zones in Southern Wyoming, J. Thorp¹⁴ has been able to trace a complete succession from grey-brown desert soils to podsols in passing from the arid plains to the relatively humid mountain regions. The grey-brown desert soils, developed under mean annual rainfalls of 5-10 inches, show the following profile characters:—

- A₁. 0-1 in. Pale grey-brown "crust and mulch" with fine vesicular structure in dry condition.
- A₂. 1-6 in. Light greyish to yellowish-brown slightly laminated soil of medium texture.
- B₁. 6-9 in. Compact cloddy, yellowish-brown soil of slightly heavier texture than above horizon. Most compact horizon in profile.
- B₂. 9-(18-40) in. Horizon of calcium carbonate accumulation of varying thickness. Calcium carbonate decreasing in lower part of horizon. White or pinkish gypsum often interspersed with upper layers of parent material below carbonate horizon.

Below 18-40 in. Parent material. Rounded gravels of various rocks. Pebbles coated with indurated calcium carbonate.

A distinct columnar structure is seen in the lower A and B horizons. Calcium carbonate is present up to the surface. In slight depressions the accumulation of calcium carbonate and gypsum becomes more pronounced.

The vegetative cover is scanty and consists mainly of sage brush, *Artemisia tridentata* and salt sage, *Eurotia lanata*.

In many parts of the arid regions of the Western United States, the calcium carbonate horizon attains a considerable

degree of induration. Under the local name of "caliche," it is frequently used as a surface for roads.

Owing to the incompleteness of leaching in soils of the pedocal class, there is a general tendency for sodium salts to occur in the profile. This is more pronounced in the more arid regions. Where, owing to topography, a ground-water level occurs near the surface, saline or alkaline soils may occur. Such soils may be considered to be analogous to the meadow and peat soils of more humid regions.

Sudan. The cotton soils of the Eastern Gezira in Sudan appear to be grey-brown semi-desert soils of the pedocal class. They are developed from wind-borne material of rather heavy texture. The rainfall is about 400 mm. per annum, of which about two-thirds falls in July and August. The natural vegetation is sparse thorn scrub.

A typical profile, described by H. Greene¹⁵, shows a surface soil of rather dark brown colour down to 2 feet. Below this is a grey layer which is penetrated by tongues of the top brown soil. In the upper part of the grey layer, ill-defined, detached lumps of grey soil occur surrounded by brown soil. Those nearer the surface contain small specks of calcium carbonate, whilst those lower down contain gypsum. At the lower limit of the grey layer, about four feet, white crumbly aggregates of calcium carbonate occur, amounting, at their greatest concentration, to 3% of the soil. Below this zone are crystals of gypsum in a matrix of yellow-brown soil.

The drying-out which occurs in the dry months of the year leads to the development of numerous cracks which extend for a considerable depth into the subsoil. These are filled with loose material blown from the surface by wind, and there is, in this way, a certain amount of circulation within the profile.

The organic matter content is low—generally below 1%, even in the dark-brown surface soil. In addition to gypsum, sodium salts are present and are closely correlated in amount with the gypsum content.

DESERT SOILS.

These form a rather ill-defined group of soils, which have not yet been sufficiently studied in detail. The actual weathering in the desert is mainly physical, but in many cases, material from more humid regions may be brought in by wind action. Owing to the absence of a closed vegetative cover, wind erosion is very marked and desert pavements occur, consisting of coarse stones or rock fragments from which the accompanying fine material has been removed by wind action. Beneath the layer of desert pavement, fine material is found.

Strongly developed crusts of calcium carbonate and gypsum are of frequent occurrence in desert regions. These may have been formed originally below the surface and subsequently laid bare by wind erosion.

Grey colours predominate in desert soils, but there are also red desert soils. In humid climates, red colours indicate degradative changes affecting the weathering complex under the influence of a lowered base status. This can scarcely be the case in red desert soils and, if the red colours have actually been developed by desert weathering, the mechanism still remains to be explained. It is possibly connected with the very high temperatures which prevail.

California. In the desert soils of Southern California, developed in colluvial fans of material of igneous origin, well marked pans are found. The degree of development of the pan is made by C. F. Shaw¹⁶ the basis of the separation of these soils into series. In the mature San Joaquin series, the profile consists of reddish-brown heavy textured soils passing through brown or reddish-brown subsoils to brown or red "iron" hardpan at 40-90 cm., with more or less calcium carbonate, overlying brown pervious unweathered material.

The profiles of San Joaquin and related soils show abundant evidence of chemical weathering, whose occurrence, in a region of scanty rainfall, may be due to natural irrigation by flood and seepage water from the adjacent mountains.

S. Africa. Red and brown soils with calcium carbonate and ferruginous pans are found in the semi-desert Karroo of South Africa.

BIBLIOGRAPHY.

- ¹ STEBUTT, A., "Die Braunerde," *Z. Pflanz. Düng.*, 1930, **15A**, pp. 134-167.
- ² KOSSOWITSCH, P., "Die Schwarzerde." *Int. Mitt. Bodenk.*, 1911, **1**, pp. 199-354.
- ³ GLINKA, K. D., *Genesis und Geographie der russischen Boden*, Petrograd, 1923, pp. 28-30; from H. STREMMER, *Grundzüge der praktischen Bodenkunde*, Berlin, 1926.
- ⁴ VON SEE, K., "Beitrag zur Kenntnis zweier Schwarzerde-vorkommen in Deutschland." *Int. Mitt. Bodenk.*, 1918, **8**, pp. 123-152.
- ⁵ HOHENSTEIN, V., "Die ostdeutsche Schwarzerde (Tshernosem) mit kurzen Bemerkungen über die ostdeutsche Braunerde." *Int. Mitt. Bodenk.*, 1919, **9**, pp. 1-31, 125-178.
- ⁶ NOVÁK, V., "Ein Beitrag zur Charakteristik der Hannaböden," *Int. Mitt. Bodenkunde*, 1923, **14**, pp. 86-106.
- ⁷ MARBUT, C. F., Private communication.
- ⁸ ANDERSON, M. S., and BYERS, H. G., "Character of the colloidal materials in the profiles of certain major soil groups." *U.S. Dept. Agr. Techn.*, Bull. 228, 1931.
- ⁹ HAWKER, H. W., "A study of the soils of the Hidalgo county, Texas, and the stages of their lime accumulation." *Soil Sci.*, 1927, **23**, pp. 475-483.
- ¹⁰ ANNETT, H. E., "The nature of the colour of black cotton soil." *Mem. Dept. Agric. India, Chem. Series*, 1910, **1**, No. 9.
- ¹¹ GRACIE, D. S., "A preliminary survey of some of the soils in Kenya." *Dept. of Agric. Kenya*, Bull. 1, 1930.
- ¹² BENNETT, H. H., and ALLISON, R. V., *The soils of Cuba*, Washington, 1928.
- ¹³ GLINKA, K. D., *Die Typen der Bodenbildung*, Berlin, 1914, pp. 133-134.
- ¹⁴ THORP, J., "The effects of vegetation and climate upon soil profiles in northern and north-eastern Wyoming." *Soil Sci.*, 1931, **32**, pp. 283-302.
- ¹⁵ GREENE, H., "Soil profile in the Eastern Gezira." *J. Agric. Sci.*, 1928, **18**, pp. 518-530.
- ¹⁶ SHAW, C. F., "The basis of classification and key to the soils of California." *Proc. 1st Int. Congr. Soil Sci.*, 1928, pp. 65-103.

CHAPTER XI.

GROUND-WATER SOILS, INCLUDING PEATS.

MEADOW SOILS AND THEIR RELATED TYPES.

THE soils described under this heading are termed " Wiesenböden " and " Gleyböden " by continental writers. In the absence of a more suitable term, the name *meadow soils* is used to describe soils whose profile characters are dominated by the occurrence of a high water table or an impervious layer impeding percolation.

These soils, which have certain affinities with the podsoles and occur under similar climatic conditions, owe their distinctive characters to the presence of a ground-water table at or near the surface. At this level, a characteristic horizon, known as a *gley* horizon, is developed. This is distinguished by the deposition of rusty streaks and mottlings of hydrated ferric oxide. The deposition of hydrated ferric oxide in the gley horizon may attain considerable proportions and form the well-known lake ore or bog iron-ore. Deposition of manganous and calcareous material may also occur, whilst there is frequently an enrichment in clay. The sub-aqueous horizons are generally grey, often with a bluish tinge, owing to the presence of vivianite. The surface horizon is grey and carries a peaty layer. The characteristic features of meadow soils can also be exhibited in situations where, above the general water table of the locality, an impervious stratum occurs which impedes drainage and gives rise to waterlogging. (Cf. pp. 56, 58, 60 *et seq.*)

Considerable variations occur in the character and thickness of the humous layer and in the development of mottling

in the gley horizon through the deposition of materials from the ground-water. The presence of a grey horizon with brown or yellowish mottlings, which are most marked in the zone of the water table, and greyish-blue sub-aqueous horizons, often with black or blue colorations, are constant indications of this soil group.

The natural vegetation of meadow soils is mainly of a *gramineous*, *juncaceous*, or *caricaceous* type. When the ground-water level approaches more nearly to the surface and permanently wet conditions prevail, the tendency is towards peat formation with the development of a peat podsol profile, consisting typically of up to 1 foot of peat underlain by grey mottled subsoil. This type passes into the peat soils.

Meadow soils are common in depressions and hollows in regions of podsolic soils. In passing from an upland to the bottom of a depression in such regions, a regular succession of profiles can frequently be observed. In the upland, with free drainage and scanty accumulation of peaty organic matter, iron podsoles may occur. In the moister lower levels, the humic layer becomes thicker and a peat podsol results. In still lower ground, the water table is sufficiently near the surface to give the gley horizon characteristic of a meadow profile, which gives place in turn to a peat profile.

Ground-water soils are essentially local in character and depend, in the main, on topographical conditions which influence the movement of underground water. Most typically they are developed in hollows, but where drainage is impeded by impervious strata on slopes, the typical characters of the gley profile may also be developed.

Ground-water soils occur throughout middle, northern, and western Europe, and indeed, throughout all the humid regions. In Europe, they attain their greatest development in the Pripet marsh region of Poland.

Analogous conditions in more arid regions may give rise to saline and alkaline soils. There are also characteristic soils associated with high ground water in the tropics. These will be considered at a later point.

Finland. B. Frosterus¹ gives the following description of a meadow soil in Finland:—

- A₁. 0-23 cm. Brownish-grey sandy peat. 0-9"
- A₂. 23-27 cm. Greyish sandy clay.
- A₃. 27-32 cm. Yellowish-brown plastic clay passing sharply into
- B. 32-45 cm. Ochreous plastic clay passing gradually into
- B-G. 45-120 cm. Grey clay with abundant rusty threads and concretions arranged in three horizontal layers at 45, 50, and 80 cm., respectively.
- G₁. (Gley horizon). 120-175 cm. Grey clay with scattered rusty threads and prismatic structure.
- G₂. 175-195 cm. Prismatic clay with superficial rusty coating on the individual columns into which the horizon is dissociated by vertical cracks.
- CG. Blue clay, occasionally inky black containing vivianite.

Germany. The soils known in parts of Germany as "Molkenböden" may be grouped with the meadow soils. K. V. Von Falckenstein² describes soils of this kind developed from the middle Bunter sandstone under conditions of impeded drainage in the Weser valley. The following profile is typical:—

- 0-3 cm. Litter layer.
- 3-16 cm. Bleached humous layer. 1-6"
- 16-50 cm. Compact layer with yellow streaks becoming more stony below.

Analyses of the successive layers show marked differences from podsoles, for there is no leaching of alkalies and alumina and only a slight downward movement of ferric oxide. The bleaching in the surface horizon is considered to be the effect of reduction of ferric to ferrous compounds.

Wales. Soils developed under conditions of impeded drainage are commonly encountered throughout Britain, but occur more frequently in districts of high rainfall.

We may distinguish two types, namely, soils in which the impedance of drainage is due to the high level of the water table, and soils in which impedance is due to the occurrence of a relatively impervious stratum in the subsoil.

As an example of the first type, a profile from near Bangor, N. Wales, may be described. The parent material

is colluvial material derived from Ordovician Shale. The soil is described for soil survey purposes as the Conway series.

The profile is as follows :—

Depth	Description	Organic Carbon	pH
0-8 in.	Grey heavy loam	5.30	6.1
8-20 in.	Grey mottled brown clay	2.47	5.8
>20 in.	Grey mottled brown heavy clay ...	1.06	5.7

The composition of the clay fraction in the successive horizons is as follows :—

Horizon	SiO ₂ %	Al ₂ O ₃ %	Fe ₂ O ₃ %	SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
0-8 in.	52.9	33.3	6.4	2.70	2.40
8-20 in.	52.9	32.9	6.3	2.73	2.43
>20 in.	49.6	32.2	8.2	2.62	2.26

This soil is typical of considerable areas of wet lowland soils in Wales. The vegetation is generally grass and rushes, together with miscellaneous water-loving plants in the wetter situations. All stages of transition to lowland peat soils may be observed.

The data for the composition of the clay fraction are instructive. In spite of the occurrence of much brown mottling in the lower horizons, there does not appear to be any considerable eluviation of ferric oxide. The low figures for this constituent throughout the profile are reminiscent of the A horizon material of a podsol profile and it may be that the whole profile consists of A horizon material removed by former erosion and deposited in its present low-lying situation.

The iron content of the clay of bottom soils is not always so low as in the profile described, and the grey colour may be partly attributed to the existence of iron in the ferrous state consequent on the reducing conditions which commonly obtain in such soils.

An example of a soil with drainage impeded by an impervious stratum is furnished by a profile from Llanfairpwll, Anglesey, mapped under the name of the Gesail series for

survey purposes. It occurs under grass vegetation. The parent material is micaceous schist boulder clay and the profile is as follows :—

Depth	Description	Organic Carbon	pH
0-10 in.	Greyish loam	3.6	6.9
10-20 in.	Greyish mottled micaceous loam ...	0.8	8.0
20-30 in.	Grey and brown mottled micaceous heavy loam	1.4	8.0

A certain amount of manganese dioxide occurs in the lower horizons, and the pH values determined by the quinhydrone method are rather too high.

The composition of the clay fraction in the successive horizons is as follows :—

Horizon	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
0-10 in.	47.6	32.0	11.3	2.50	2.05
10-20 in.	45.8	29.6	15.0	2.63	1.99
20-30 in.	45.6	29.2	14.6	2.66	2.02

There is a certain eluviation of ferric oxide, but there is no very decided difference between the composition of the clay fraction in the different horizons.

In the two profiles described above, the mottled sub-surface horizons may be equated with the gley horizons of continental writers. It should be noted that the brown or reddish-brown appearance is not due to a deposition throughout the mass of the soil as in the B horizon of podsol profiles, but rather to superficial deposition on the structural elements. This is reflected in the comparatively slight enrichment in ferric oxide shown by the analytical figures for the clay fraction.

One feature of meadow soil profiles is the fairly sharp decrease in organic matter content in passing from the turf layer. The moisture conditions are such as to restrict root development to the surface, so that comparatively little organic matter is present in the layers subjected to water-logging.

England. Certain of the soils of central and eastern England derived from heavy clay formations might be appropriately grouped with the meadow soils. The following profile at Lolworth, Cambridgeshire, on Kimmeridge Clay is described by F. Hanley³ :—

- 0-9 in. Dark brown clay loam with flints: mottlings.
- 9-21 in. Grey brown clay with light brown mottlings.
- 21-40 in. Lead-grey clay with brown mottlings.
- >40 in. Lead-grey clay; yellow seams in upper portion and pockets of crystalline gypsum.

The analytical figures show an almost complete leaching of carbonates and a marked accumulation of gypsum below 40 inches. The soil contains 32.8% clay, but the lower horizons have from 60-70% clay.

VLEI* SOILS.

This name has been given to certain African soils which occur in depressed or basin-shaped areas subject to seasonal wetness. They may be black, brown, or grey in colour and often show mottling due to the prevalence at certain seasons of waterlogged conditions. Ironstone concretions (murrum) are frequently present, and calcium carbonate nodules are also encountered.

Kenya. D. S. Gracie⁴ gives the following description of a profile in the South Thomson's Falls district of Kenya :—

- 0-9 in. White soil with mottling. Poor accumulation of organic matter at the surface. A band of iron concretions separating this layer from the next.
- 9-31 in. Compacted organic clay layer, whitish at first, but shading into black and becoming brown at its lower limit.
- 31-60 in. Yellowish material, soft and friable in the first six inches, then becoming harder. Large calcium carbonate concretions present.

The surface soil has a pH of 4.11 and contains 0.183% nitrogen.

*This Afrikaans word, pronounced *flay*, may be conveniently used for the tropical and sub-tropical analogues of the meadow soils.

Tanganyika. G. Milne⁵ describes a typical vleï soil in a saucer-shaped depression at Muhesa, Tanganyika. The altitude is 600 feet and the rainfall 50 inches, with maxima in March-May and November-December. The original vegetation was tall dense grass. The profile is:—

0-4 in. Friable grey-black surface soil.

4-10 in. Inky black sticky subsoil.

Below 10 in. Mottled yellow soapy clay to 11 feet.

There are many ferruginous concretions through the subsoil and traces of calcium carbonate from 18 inches downwards. The following data for the Muhesa vleï soil were obtained in the writer's laboratory (Table XXIX).

TABLE XXIX. — ANALYTICAL DATA FOR MUHESA VLEI PROFILE.

Depth	Clay %	Organic Carbon %	pH	Clay Fraction			
				SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /R ₂ O ₃
0-4 in.	27.6	2.99	7.6	48.0	20.5	19.8	2.47
4-8 in.	31.9	2.56	6.9	47.1	21.2	20.0	2.39
26-32 in.	45.5	0.85	6.3	43.8	34.7	10.8	1.80

Cuba. An example from Cuba of what would appear to be an analogue of the vleï soils is furnished by the Maboa series of H. H. Bennett and R. V. Allison⁶. These soils are developed on outwash from siliceous rocks in regions of flat topography with poor drainage. The profile consists of:—

0-3 in. Grey silt loam or fine sandy loam with some rusty brown mottling. Acid reaction.

3-8 in. Friable yellowish-grey or mottled pale yellow and grey material.

8-12 in. Pale yellow and friable material with abundance of perdigón (concretions).

>12 in. At less than two feet. Mottled red and bluish-grey plastic impervious heavy clay.

The Coxville series of Cuba is similar to the Maboa series, but concretions are comparatively rare.

San Domingo. Certain soils in San Domingo described by E. Balzarotti⁷ as podsolised may owe their character to ground-water effects. They occur under savanna.

S. Rhodesia. H. B. Maufe⁸ describes a vlel soil derived from dolerite at Salisbury, S. Rhodesia. The profile consists of :—

- 0-9 in. Black soil.
- 9-28 in. Black clay subsoil.
- 28-43 in. Yellowish-brown clay with rounded blocks of fresh dolerite.

The top soil is coarsely granular, but the subsoil has a columnar or prismatic structure. Although the pit in which this profile was seen was waterless at the end of the dry season, the bottom part was distinctly wet, indicating poor drainage conditions. In a better drained situation in the immediate locality, the same rock gives rise to a red soil.

BLACK PRAIRIE SOILS.

Many of the black prairie soils of Canada and the United States appear to be essentially meadow soils. In a description of Saskatchewan soil profiles, A. H. Joel⁹ mentions a type of "black park land" profile, consisting of a "black or almost black A horizon of mixed loose and fine granular structure; a B horizon of some shade of brown, heavier and more compact than the surface, frequently somewhat columnar in structure and usually fine to medium granular when crumbled; a B₂ horizon of grey to light grey colour, frequently with a yellowish or brownish tinge and flowing to granular structure (zone of carbonate accumulation); and a C horizon of grey to dark grey colour mottled with flecks or stains of reddish-brown ferric oxide and frequently with concretions or splotches of calcium carbonate, calcium sulphate or both. The C₂ horizon is usually a more uniform dark grey or somewhat bluish dark-grey." A shallow and a deep phase are distinguished, the deep phase occupying lower elevations. Whilst the shallower phase is considered to be well drained and may be a type of tshernosem, the deep phase shows evidences of poor drainage conditions in iron stains and mottlings. The black layer varies from about 3 to 8 inches in the deep phases. It is thus rather shallower

than in typical tshernosems to which, on the grounds of carbonate accumulation, these soils might be assigned.

From this circumstance and the analytical evidence of mechanical and chemical eluviation, Joel excludes them from the tshernosem group. It would appear that the deeper phases, at least, must be assigned to the meadow soil group.

Black soils in Alberta, described by F. A. Wyatt and J. D. Newton¹⁰, show a somewhat similar profile, with an organic layer of about 12 inches in thickness overlying about 30 inches of grey columnar to granular clay loam or clay, with carbonate accumulation in the lower 6 inches. This is underlain by light yellow clay with occasional iron stains. The data for the nitrogen content of successive layers, both of the Saskatchewan and the Alberta black soil profiles, show a sharp decrease in organic matter in passing from the dark layer. This is markedly different from the gradual decrease in organic matter with depth observed in the tshernosems and suggests the effect of impeded drainage conditions.

Considerable discussion has centred round the so-called prairie soils of the Middle-West of the United States. Whilst many of the European workers assign them to the meadow soils, C. F. Marbut¹¹, whilst admitting that, in the earliest stage of topographical development, impeded drainage may have obtained, considers the present drainage to be unimpeded. There are, however, areas in South-Central Iowa, North-Eastern Missouri, and parts of Southern Illinois where meadow soils occur owing to the survival of drainage impedance.

China. C. F. Shaw¹² refers to a class, known as Sajong soils, occurring in the central plains of China. They occur with an almost level topography. They show horizons of calcium carbonate accumulation at depths varying from one to six feet. From the description of the drainage conditions, it seems likely that they belong to the meadow soils rather than to the tshernosem group. The calcium carbonate would thus represent a deposit in the vicinity of a water table.

CONDITIONS OF DEVELOPMENT OF GROUND-WATER SOILS.

In the present chapter, we have discussed soils with impeded drainage from regions with widely differing climatic conditions. Whilst they occur more frequently in humid regions such as North-West Europe, they may be found also under comparatively dry climates such as that of the Canadian prairies. The common characters which distinguish them are greyish to black surface horizons sharply defined from the sub-surface horizons which, by their iron stains, mottlings, or concretions, and also, in many cases, by the presence of gypseous or calcareous horizons, give evidence of the alternation of aerobic and anaerobic conditions due to variable ground-water levels. The deposition of hydrated ferric oxide may often reach notable proportions, as in the soft bog iron-ore of northern lands and the harder "murram" or "mocarrero" of the tropics.

Soils developed under conditions of impeded drainage may, in some cases, show resemblance to the black earths. The relatively high content of organic matter often developed under neutral or alkaline conditions may result in a dark colour in the surface horizons. Further, the drainage impedence and the high base status both tend to the development of a siliceous type of weathering complex analogous to that of the tshernosem group. Finally, the presence of a ground-water table may in many cases be associated with the deposition of calcium carbonate in the adjacent horizons.

Meadow and vlei soils are, however, distinguished from the tshernosems by the absence of the granular structure and by the occurrence of rusty mottlings or streaks, indicating the alternation of oxidative and reductive conditions.

In the soils of this group examined by the writer, the mineral colour of the surface horizons is generally grey and never red or reddish-brown. Such analyses as have been made show the clay complex to be of a siliceous type with silica-sesquioxide ratios in excess of 2.0. It would be of interest to determine whether this is a constant feature.

S. AFRICAN BLACK TURF SOILS.

There is a class of soils occurring typically in the Transvaal, but also found with no essential difference in Southern Rhodesia, to which the name *black turf* has been applied. The profile consists of a variable depth, commonly about three feet, of grey-black or brownish-black clay, granular when dry but waxy when wet. Calcareous concretions occur in the lower part of the layer, increasing in number to form eventually a friable calcareous layer, below which there is a transition to decomposing norite rock. In some cases the calcareous layer is absent and the black layer rests directly on the decomposing rock.

B. de C. Marchand,¹³ in a discussion of the origin of these soils, controverts the theory advanced by C. F. Marbut,¹⁴ that the black turf soils belong to the tshernosem group, as might be expected from their black colour and the presence of a layer of carbonate accumulation. He shows that their occurrence is strictly associated with the outcrop of norite rock, whilst, under identical climatic and topographical conditions, diabase yields a completely leached friable red loam. They cannot be described as vleis soils since they are not limited to areas of depressed topography. The critical factors in their formation appear to be impeded percolation, consequent on the impervious character of the clay formed by the weathering of the norite, and their high base status, which results in the development, as elsewhere in hot climates, of a dark type of organic matter. The dark colour of these soils may convey an erroneous impression of their organic matter content, which is not generally higher, and may in some cases be actually lower, than that of the adjoining red loams.

The clay fraction of the black turf soils reflects the restriction of percolation, for it is markedly siliceous in character. Whilst the clay fraction of the leached friable red loams has a silica-sesquioxide ratio of 2.0 or less, the clay fraction of the black turfs has a silica-sesquioxide ratio of 3.0 or more.

PEAT SOILS.

Peat soils are distinguished by the high proportions of organic matter which they contain. They are soils in which the residues of natural vegetation are humified rather than oxidised. A distinction is sometimes drawn between peat soils and mineral soils, but it is somewhat difficult to assign a limiting content of organic matter above which a soil is to be considered a peat. Whilst 15% of organic matter may suffice to give a pronounced peaty character to a light sand, a heavier soil may contain up to 20% of organic matter and still retain the appearance and character of a mineral soil. Typical peats, however, generally contain at least 50%, and may in some cases consist almost entirely of organic matter.

The humification which results in the formation of peat is, except in the case of heath peat and coniferous forest peat, of an anaerobic character, and takes place either under water, as in the formation of fen peat, or in situations where water-logging and consequent exclusion of air is the rule. From this it follows that peat is most readily developed under cool humid climates. But whilst they are commonly found under such conditions, peats can occur in the tropics, as for example, in mangrove swamps. Notable deposits of peat occur under the sub-tropical conditions in Florida.

Typically, peats are formations of temperate and cool climates. They are more widespread in northern regions, partly because of the extensive spreads of glacial drift with its associated lakes, and partly because the scanty evaporation associated with low temperatures results in a greater proportion of the rainfall being available as soil-water.

Most of the peat of Northern Europe and America has been formed in basins formerly occupied by lakes. Existing deposits represent different stages in a process which has proceeded since the close of the Glacial Period and is still in progress. The relationships of the different types of peat to each other will be evident from a description of this process.

The first stage of peat formation takes place under lacustrine conditions. In the shallow water near the shore,

associations of such plants as *Phragmites*, *Carices*, *Juncus spp.*, and *Equisetaceæ* become established. The residues of these plants accumulate at the bottom, where they are mixed to some extent with lake mud and undergo anaerobic decomposition. The littoral association encroaches more and more on the free water surface until the lake is completely filled up with the humified residues of the plants which have inhabited it, mixed with a certain proportion of mineral matter carried by the stream feeding the lake and deposited as sediment. Examples of all stages of this process may be seen in any region of shallow lakes. The peat thus formed is known as low-moor or fen peat ("Niederungsmoor" or "Flachmoor").

With the disappearance of the free water surface and the accumulation of humified material, a gradual change in the type of vegetation sets in: mosses (e.g., *Hypnum*), *Calluna vulgaris*, *Erica tetralix*, *Molinia cærulea*, and finally trees, such as alders (*Alnus*), birches (*Betula nana* and *pubescens*) and pines (*P. sylvestris* and *uliginosus*) make their appearance, whilst the earlier lacustrine plants die out. The peat accumulated from the residues of the new vegetation is now of a different type and may be recognised, when this stage is past, by the presence of tree remains.

The conditions at the close of the fen stage and before the establishment of tree growth are relatively dry, owing to the growth of the peat deposit above the original water level. The establishment of tree growth, however, by checking evaporation, results in moister conditions, and a further stage in development is marked by the growth of sphagnum moss. Meanwhile, there has been a progressive decline in plant nutrient status. The plants which contribute to fen peat are relatively rich in plant nutrients and grow in water containing considerable proportions of mineral matter in solution and suspension. As the peat lies above the fen level and contributions from telluric waters are diminished, the type of vegetation which prevails consists of plants which make scanty demands for nutrients from the medium on which

they grow. The peat formed from their decay becomes progressively poorer in nitrogen and mineral constituents. This, and the wetness consequent on the high retentivity of the sphagnum for moisture, eventually results in the annihilation of tree growth, and the peat deposit now enters on its final phase as an upland or high-moor peat or bog (Hoch-moor) in which the principal plants are sphagnum species and cotton grass (*Eriophorum*).

In a fully developed profile, a section will disclose a complete succession from the beginning of the fen peat stage. W. Bersch¹⁵ gives the following typical example:—

1. Vegetation layer.
2. Younger moss peat, 140 cm.
3. Light coloured transition layer (heath), 20 cm.
4. Older moss peat, 170 cm.
5. Woody peat, 85 cm.
6. Fen peat (*Phragmites* and *Carices*), 70 cm.
7. Peat mud, 15 cm.
8. Mineral horizons.

Horizons 1 to 4 represent the high level peat; horizon 5, the transition (woodland) stage; and horizons 6 and 7, the fen peat. The transition layer between the older and the newer peat moss is generally observed in continental high level peats and is considered to represent a period of relatively dry conditions involving an intermission of sphagnum peat formation, and the establishment of heath.

Peat deposits may reach to a depth of 50 feet or even more. The rate of accumulation has been variously estimated. Whilst the vertical rise of the surface may amount to 1-2 cm. per annum, the actual growth of solid peat must be considerably less. An estimate may be gathered from an example at Laibach, given by Bersch, where a Roman road of about the first century A.D. was found below 120 cm. of peat. Assuming 1,800 years as the upper limit of time, the rate of accumulation is about 0.7 mm. per annum.

Lacustrine peat is markedly different in character and composition from sphagnum peat, for not only are the plants from whose remains it has been formed richer in ash con-

stituents, but there is also an admixture of sedimentary material of terrestrial origin.

S. A. Waksman and K. R. Stevens¹⁶ examined samples of lacustrine and sphagnum peats and their results shown in Tables XXX and XXXI show clearly the contrast in composition between the two types.

TABLE XXX. — CHEMICAL COMPOSITION OF SEVERAL HORIZONS OF A LOW-MOOR PEAT PROFILE FROM NEWTON, NEW JERSEY, U.S.A., ON BASIS OF DRY MATTER (WAKSMAN AND STEVENS).

Horizon	Ether Soluble Fraction	Water Soluble Fraction	Hemi-Celluloses	Cellulose	Lignin	Crude Protein	Ash
A ₁	0.66	3.08	10.31	0	38.35	22.48	13.22
A ₂	1.10	1.24	8.95	0	50.33	18.72	10.13
A ₃	0.49	2.31	7.02	0	57.83	14.81	10.15
A ₄	0.78	1.14	7.51	0	42.10	19.81	15.00
Lake peat ...	0.67	0.81	12.14	0	33.25	19.38	24.87
Bottom mud	0.36	1.24	5.92	0	15.62	9.81	59.55

TABLE XXXI. — CHEMICAL COMPOSITION OF SEVERAL HORIZONS OF A HIGH-MOOR (SPHAGNUM) PEAT PROFILE FROM OLDENBURG, GERMANY, ON BASIS OF DRY MATTER (WAKSMAN AND STEVENS)

Horizon	Ether Soluble Fraction	Alcohol Soluble Fraction	Hemi-Celluloses	Cellulose	Lignin	Crude Protein	Ash
Younger sphagnum 1	3.08	...	16.88	19.44	34.04	5.23	1.72
" " 2	6.12	...	11.09	13.62	49.56	5.11	1.85
" " 3	3.96	...	16.24	19.91	38.26	6.58	1.50
" " 4	5.02	...	15.44	16.38	40.72	6.11	2.36
" " 5	4.99	...	15.12	17.14	43.82	5.10	1.82
Heath peat ...	7.60	...	8.44	9.76	54.38	7.00	1.71
Older sphagnum 1 ...	5.73	...	9.08	12.38	52.50	5.78	1.38
" " 2 ...	5.18	3.87	9.48	9.63	53.30	6.20	1.13
" " 3 ...	4.28	3.52	10.86	13.74	45.21	4.26	1.16
" " 4 ...	5.17	3.72	10.82	10.42	52.62	5.63	2.22
Woody peat ...	5.69	5.13	4.91	4.08	59.86	9.08	5.03

The first point to be noticed is the contrast in the ash contents. Whilst, in the low-moor peat, the ash varies from about 10 to 25%, in the high-moor, the ash is between 1 and 2%, rising in the woody peat, which represents a transition to low-moor, to over 5%.

Secondly, whilst cellulose is completely decomposed under the conditions obtaining in low-moor, the high-moor peats contain between 10 and 20% of this constituent. The transition to low-moor is shown by the fall in cellulose to 4.08% in the woody peat.

Thirdly, the crude protein content of the low-moor peats, excluding the bottom mud, varies from 18.72 to 22.48%, whilst in the high moor the range is from 4.26 to 7.00%, the woody peat indicating the transition with 9.08% of crude protein. The highest figure in the high-moor peat, 7.00%, is found in the heath peat, the so-called "grenzhorizont," which is a constant feature of continental high moors and may indicate a dry period.

W. L. Davies,¹⁷ in a comparison of the nitrogenous matter of peats, found evidence of more rapid protein degradation under fen conditions than in upland peat or heath conditions.

The complete succession from lowland peat, through forest peat, to upland peat can only be observed in basins or depressions formerly occupied by lakes. In other cases, development may begin at the forest peat or the upland peat stage.

MOUNTAIN PEAT.

There is a type of peat, termed by continental writers "Hangmoor," which is of common occurrence in mountainous regions. It is formed along slopes down which water is continually percolating from springs in higher ground. The character of the peat will depend on the nature of the percolating waters. Where these are poor in dissolved material, for example, surface water, or waters from highly acid rocks, the peat will be of the high-moor or

sphagnum type. With richer waters, on the other hand, peats of the fen type may be developed, whilst in some cases trees may be established. Extensive deposits of acid peat of the upland type are found directly on boulder clay or colluvial material throughout the mountainous regions of Great Britain.

HEATH PEAT.

This type of peat is developed under conditions of relatively free drainage and recurrent drought. The vegetation consists of such plants as *Erica cinerea*, *Calluna vulgaris*, *Vaccinium oxycoccus*, *Festuca* species, and certain mosses. The peaty layer is generally less than a foot in thickness and is very acid in character. The mineral soils below show successively the bleached horizon and the illuvial horizons of the podsol profile.

FOREST PEAT.

Under coniferous forest, there is generally a layer of peaty humus formed from leaf litter. Here, also, as in the case of heath peat, the mineral horizons of the podsol profile are found beneath.

BIBLIOGRAPHY.

- ¹ FROSTERUS, B., "Beitrag zur Kenntris der Bodenbildung in Tonen der humiden Geigenden." *Int. Mitt. Bodenk.*, 1914, **4**, pp. 105-137.
- ² VON FALCKENSTEIN, K. V., "Die Molkenböden des Bram-Reinhardswaldes im Buntsandsteingebiet der Oberweser." *Int. Mitt. Bodenk.*, 1914, **4**, 105-137; 1915, **5**, 77-101.
- ³ HANLEY, F., Private communication.
- ⁴ GRACIE, D. S., "A preliminary survey of some of the soils in Kenya." *Dept. of Agric. Kenya, Bull.* 1, 1930.
- ⁵ MILNE, G., Private communication.
- ⁶ BENNETT, H. H., and ALLISON, R. V., *The Soils of Cuba*, Washington, 1928.
- ⁷ BALZAROTTI, E., "Über das Auftreten von Boden mit Podsolflecken in Santa Domingo." *Soil Res.*, 1930, **2**, pp. 1-10.
- ⁸ MAUFE, H. B., "On the formation of red soil and of black vlei soil from dolerite at Salisbury, Southern Rhodesia." *S. Afr. J. Sci.*, 1928, **25**, pp. 156-167.

- ⁹ JOEL, A. H., "Predominant Saskatchewan soil profiles correlated with soil development factors in northern latitudes." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 200-222.
- ¹⁰ WYATT, F. A., and NEWTON, J. D., "Alberta soil profiles." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 358-366.
- ¹¹ MARBUT, C. F., "Transcontinental Excursion Guide." *1st Int. Congr. Soil Sci.*, 1928, p. 110.
- ¹² SHAW, C. F., "Soils of China." *Geol. Survey of China*, 1930, Bull. 1, pp. 1-38.
- ¹³ MARCHAND, B. DE C., "The origin of black turf soils in the Transvaal." *S. Afr. J. Sci.*, 1924, **21**, pp. 162-181.
- ¹⁴ MARBUT, C. F., in H. L. SHANTZ and C. F. MARBUT. *The vegetation and soils of Africa*, New York, 1923.
- ¹⁵ BERSCH, W., *Handbuch der Moorkultur*, W. Frick, Vienna and Leipzig, 1912.
- ¹⁶ WAKSMAN, S. A., and STEVENS, K. R., "Contribution to the chemical composition of peat." II.—Chemical composition of various peat profiles. *Soil Sci.*, 1928, **26**, pp. 239-251.
- ¹⁷ DAVIES, W. L., "The proteins of different types of peat soils." *J. Agric. Sci.*, 1928, **18**, pp. 682-690.

Note—T. Rigg (New Zealand J. Sci. Techn., 1929, **11**, 231-241) describes certain soils, termed Pakihi soils, occurring in the Nelson Province of New Zealand, in which the profile consists, typically, of humous peaty sand or silt, succeeded by brown and grey silt which passes into cemented sand and gravel. They are considered to be podzols, but the prevalence of drainage impedance and the character of the native vegetation suggest that they would be more appropriately grouped with the ground-water soils.

CHAPTER XII.

SALINE, ALKALINE, AND SOLOTI SOILS.

CONDITIONS OF OCCURRENCE.

THE soils to be described in the present chapter occur most commonly, but not exclusively, under arid climates. They owe their distinctive characters either to the presence of an actual excess of sodium salts, or to the predominance of sodium among the exchangeable bases—the latter a consequence of the former presence in the soil of sodium salts. In some cases, potassium salts may be present in appreciable amounts, but it is unusual for soils to occur in which potassium is the dominant exchangeable base.

The most commonly occurring groups are the saline, white alkali or *solontshak* soils, and the alkaline, black alkali, or *solonetz* soils. To these must be added a third group which has been brought to the notice of soil students by the Russian pedologists, namely, the *soloti* or degraded alkali soils.

Before entering on a description of some representative types of these groups, it will be necessary to enquire into the conditions under which saline, alkaline, and soloti soils occur.

The first pre-requisite for their occurrence is the presence of sodium salts in the soil or the parent material. These originate usually from underground water in situations where the water table is relatively near the surface. Such conditions occur generally in areas of basin shaped topography. The salts may have originated directly from the chemical weathering of silicates, resulting in the accumulation in the underground water of the soluble products of

hydrolysis, or they may represent the vestiges of former seas or salt lakes. Stated generally, the most usual mode of origin is from saline telluric waters either in the vicinity of inland seas or salt lakes, or in depressions where the water table is at or near the surface. In such cases there has usually been a change in the hydrological conditions. In primary saline soils, there has been a lowering of the water surface whereby soils or their antecedent parent materials have emerged. In this way, for example, the saline soils of the Great Basin in the United States have originated, for they occupy an area formerly covered by the ancient Lake Bonneville, represented to-day by the Great Salt Lake. The fall in level has doubtless been accompanied by a concentration of soluble salts in the remaining water. A similar fall in water level is also postulated in the Caspian Basin.

In secondary or regraded saline soils, the salification has taken place by a rise in the level of salt-bearing ground-water. This may have taken place naturally, as, possibly, in parts of Hungary; but more frequently it is encountered as a result of human interference. W. W. Mackie¹ records that in 1873 the water table at Fresno, in the San Joaquin valley of California, was 65 feet below the surface and the soil was practically free of salts. As a result of irrigation without adequate under-drainage, the level of the underground water in 1888 was only 2-3 feet below the surface and the soil was definitely affected with "alkali." Similar instances can be adduced from most irrigated areas where large tracts of formerly fertile soil have been ruined through the rise of ground-water.

We may class with the secondary saline and alkaline soils, those which, without any marked change in water table, have undergone change through the use of irrigation waters in which sodium is the predominant cation. The continued use of such waters leads to a gradual replacement of exchangeable calcium, by sodium, so that, even although no considerable proportion of sodium salts is present, the soil becomes a sodium soil.

SALINE SOILS.

These soils are the white alkali soils of the earlier American writers, and the *solontshak* soils of the Russian school. They contain an excess of sodium salts, generally the chloride or sulphate, and ~~have~~ a flocculated structure. They occur usually in depressions, and the distribution of salts varies with the season. During drought they show white efflorescences of sodium chloride or sulphate, from which the American term white alkali is derived. During rain, the salts deposited at the surface are dissolved and temporarily washed down towards the water table. A considerable variety of possibilities presents itself; for the character of the soil will depend on the position of the water table, the amount and distribution of the rainfall, the concentration and composition of the soluble salts, and the general character of the soil of the region. Saline soils occur most commonly in arid, semi-arid, and semi-humid regions and may represent modified tshernosems, chestnut earths, or grey earths. Where the water table lies so near to the surface as to maintain permanently moist conditions, saline peaty or saline meadow soils may result.

Russia. Saline soils are generally described by Russian writers as structureless in contradistinction to the structural alkaline (solonetz) soils. A typical profile from the vicinity of Akmolinsk is described by G. Tumin² :—

- A₁. 0-1 cm. Light grey salt crust with carbonates, passing to a dark grey looser structureless horizon. Gradual transition to
- A₂. Brown with grey streaks and mottlings. Weak carbonate reaction. Numerous small flecks of salt to 30 cm. From 30-45 cm., larger flecks.
- C. Light brown salt bearing loam, containing carbonates.

The whole profile is moist and the saline ground-water is encountered at 110 cm.

Glinka³ describes a saline soil with affinities to the chestnut earths in the Jenisseisk province. The profile is as follows :—

- 0-1.5 cm. Loose dusty soil with plant roots.
 1.5-9 cm. Rather more compact chestnut soil without marked structure.
 9-19 cm. Lighter coloured than above horizon but especially porous.
 19-44 cm. Lighter coloured and porous.

Finland. Saline soils may occur in humid regions where ground-waters containing salts may occur. B. Aarnio⁴ describes certain Finnish soils in which saline ground-water occurs. The following table shows the composition of the salts in different horizons of a typical profile consisting of :—

- 0-10 cm. Dark brown peat.
 10-20 cm. Fine sand.
 >20 cm. Litorina clay.

TABLE XXXII. — SALTS IN VEGETATION-FREE SALINE SOIL, ISOKYRO, FINLAND (AARNIO).

Depth				0-10 cm.	10-20 cm.	25-35 cm.
Percentage Water-soluble Salts ...				2.45	0.20	0.22
SiO ₂	...	...		0.29	—	12.26
Al ₂ O ₃	...	...		10.77	7.32	3.66
Fe ₂ O ₃	...	...		0.31	—	—
CaO	...	...		23.77	13.24	12.44
MgO	...	...		1.65	6.92	10.89
K ₂ O	...	...		0.41	—	1.46
Na ₂ O	...	...		4.16	16.42	7.32
SO ₃	...	...		56.78	48.84	48.76
Cl	...	...		0.85	7.12	3.20
Total ...				99.99	98.86	99.99

At 15 cm., there is a gley horizon showing brown streaks in the sand, whilst the clay is coloured yellowish-brown. The reaction of the surface horizon is less than pH 4. Removal of salts by drainage involves a lowering of

the water table, which is more than 1 metre below the surface. Capillary rise of salts is inappreciable, a circumstance which is connected with the humid climate. Saline water at this depth in an arid climate would dominate the character of the soil.

Saline soils, since they often occur in situations where the water table is comparatively near the surface, may be expected to show affinities with the meadow soils. This is shown in some profiles by the presence of rusty flecks or mottlings in the lower horizons.

In extreme cases, saline soils are devoid of vegetation, as for example, in the vicinity of the Great Salt Lake, and the Dead Sea. In other cases, there is an open association of halophytic plants such as salt bush. With lower contents of soluble salts there is a transition to the normal soil of the region, which may nevertheless exhibit the characteristic structural and chemical features of the saline group.

ALKALINE SOILS.

These soils, which are closely associated with saline soils, are termed black alkali by the American workers and *solonetz* by the Russians. They are characterised by the presence of sodium carbonate. Whilst the saline soils are in a state of flocculation, the alkaline soils have a strongly alkaline reaction and are deflocculated. The presence of alkaline soils in an area of saline soils is marked by depressions corresponding with the decrease in pore space implied by the change of structure. In these depressions, during rain, alkaline solutions of humic matter accumulate and, on drying, leave black deposits which give to these soils the American name "black alkali."

The alkaline soils are further distinguished by the development of a structure profile. Typically, this consists of a laminated superficial horizon overlying a deep layer showing a columnar (rounded tops) or prismatic structure. The rounded tops of the columnar elements and the vertical

sides frequently carry a white mealy coating of salts. The soil colours are grey or greyish-brown throughout.

The "structure" of alkaline soils is a consequence of their deflocculated state. When wet, the whole body of the profile consists of structureless material in the single grain state. On drying, shrinkage occurs and prismatic or columnar structures develop.

Ukraine. The following data are cited by D. G. Vilensky⁵ for a profile in the Dnieper region:—

- 0-19 cm. Light grey, platy, fine sandy eluvial horizon. The surface of the plates is powdered.
- 19-29 cm. Illuvial, dark grey, columnar horizon (columns 8-11 cm.). The columns have rounded tops with a meal-like efflorescence which also appears in places on the vertical sides of the columns.
- 29-48 cm. Transitional prismatic.
- 48-78 cm. Accumulation of sulphates. Light loess-loam with minute inter-layers of gypsum.

Another profile described by Vilensky showed the following characteristics:—

- 0-1.5 cm. Humus horizon. Semi-peaty, grey-brown loam.
- 1.5-10 cm. Dark brown short columnar. Columns are 4 cm. in diameter, clayey, compact, and do not effervesce with acid.
- 10-27 cm. Rather lighter coloured, breaking into large pillars, heavy loam.
- 27-83 cm. Dark brown with grey spots of effervescent salts. Compact loamy.
- 83-100 cm. Muddy brown with very numerous white spots. Loose non-structural.
- 100-120 cm. Bluish-grey flowing sand with water at 120 cm. under pressure.

The above profile contains chloride and sulphate in the upper horizons.

Hungary. The Hungarian alkali soils (Szík lands) contain very small proportions (0.15-0.20%) of soluble salts. They exhibit impervious horizons, considered by A. A. J. De Sigmond⁶ essential to the formation of such soils. The contrast between alkali and non-alkali soils in the character of the exchangeable cations is shown by the data in Table XXXIII.

TABLE XXXIII. — EXCHANGEABLE BASES IN ALKALINE AND NON-ALKALINE SOILS (DE SIGMOND).

Soil		Milli-equivalents per cent.				
		CaO	MgO	K ₂ O	Na ₂ O	Total
Bad Alkali	{ Hortobàgy 1	2.60	1.34	0.94	9.26	14.50
	{ Bèkèscsaba 5	8.32	3.86	0.14	7.30	19.62
	{ Mezohègyes	17.65	2.54	0.55	4.68	25.42
Non-alkali	{ Keszthely 1	15.96	1.75	0.33	0.19	18.23
	{ Keszthely 22	15.18	1.82	0.41	0.12	17.53

Whilst, in the bad alkali soils, the proportions of Na₂O to the total bases are 66.35, 37.21, and 18.41%, respectively, in the non-alkali soils, the proportions are 1.01 and 0.41%, respectively.

RELATIONSHIPS OF SALINE TO ALKALINE SOILS.

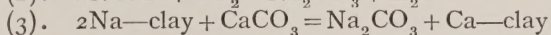
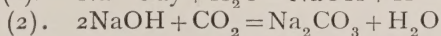
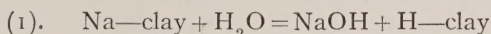
The relationship of alkaline to saline soils is now intelligible owing to the investigations of Gedroiz, De Sigmond, Kelly, and others, on the base exchange reactions of these soils. The presence of an excess of sodium salts in saline soils results in the partial or complete replacement of their exchangeable calcium by exchangeable sodium. So long as excess of sodium salts is present they remain in the flocculated condition with neutral or slightly alkaline reaction. With removal of excess of sodium the reaction becomes markedly alkaline and deflocculation occurs. The process has been already discussed in Chapter V.

There is no hard and fast line of division between saline and alkaline soils, and all stages in the transition between the two types may be encountered. This, in addition to the great variety of possibilities in the conditions antecedent to salification, results in a very great variation in the character of these soils. The principal factors are, however, the position of the water table, the composition of the ground water, the nature of the soil and the degree of leaching, either naturally by rainfall or artificially by irrigation.

The problem of alkaline and saline soils is of the greatest importance for irrigated agriculture. Irrigation is usually practised in valleys and, unless special measures are taken to ensure drainage, the ground-water level rises, bringing sodium salts within the effective capillary system of the surface horizons. This leads in time to the replacement of calcium by sodium in the absorbing complex and, in addition, may involve such an increase in salt concentration in the soil moisture as to inhibit plant growth. By the leaching action of irrigation water, the saline soil may become changed into an alkaline soil with deflocculation and increase in alkalinity.

In some cases the salification may result from the irrigation water itself having excess of sodium over calcium ions, as is so frequently the case in S. Africa. The soil becomes changed from the calcium to the sodium soil with consequent deterioration in fertility.

It should be remarked that, although the presence of sodium carbonate is a characteristic of alkaline soils, the development of their distinctive properties is not dependent on the occurrence of large proportions of this salt. If, by the continued action of sodium salts, a soil is brought to the state in which the exchangeable calcium is replaced by sodium, the resulting sodium soil is potentially an alkaline soil and becomes deflocculated as soon as the concentration of the excess salts falls sufficiently low to permit the hydrolysis of the sodium clay. Under these conditions, sodium carbonate may be formed either by the reaction of the sodium clay with calcium carbonate, or by the action of carbon dioxide on the sodium hydroxide liberated by hydrolysis. These reactions may be represented by the equations:—



Sodium carbonate should thus be regarded as one of the consequences of alkalisation rather than as the cause

of the distinctive characters of alkaline soils. The percentage of sodium carbonate is in itself no measure of the degree of alkalinisation of a soil. If carbon dioxide be excluded, it is possible to prepare completely deflocculated soils from which sodium carbonate is entirely absent.

An example of the artificial alkalinisation of a soil by the action of sodium salts is seen in certain manurial plots at the Rothamsted station which have received repeated heavy dressings of sodium nitrate. The soil has now become sticky and impervious, and exhibits some typical features of an alkaline soil.

SOLOTI OR DEGRADED ALKALI SOILS.

The hydrolysis of sodium soil with production of sodium hydroxide involves also the formation of the hydrogen soil, as shown above. In the presence of excess of calcium carbonate, calcium soil is speedily reformed, if indeed it is not formed by the direct inter-action of the sodium soil and calcium carbonate. In the absence of calcium carbonate, however, degradation of the hydrogen soil into silicic acid and sesquioxides occurs. The sesquioxides become leached out, leaving a bleached eluviated horizon relatively rich in silica. The process is in its results similar to podsolisation, but the actual degree of destruction is greater. The A horizon of *soloti* soils may contain as much as 12% of free silica of secondary origin.

A typical soloti profile is described in the following terms by Korotky⁷ :—

- A₀. Peaty layer with grass cover; 3-5 cm.
- A₁. Grey or ashen-grey layer with tongue-like projections from the peat layer; 1-10 cm.
- A₂. Whitish grey, with numerous rust spots, and ferruginous concretions; typical podsollic laminated structure; 10-15 cm.
- B₁. Brown varying to rusty or bluish-grey (humus); abundant humous mottlings; a sticky clay; numerous "ortstein" grains in the upper portion, forming in the lower portion a black or dark brown material (iron and manganese compounds); on drying breaks into prismatic lumps.

Saline, alkaline, and soloti soils generally occur in association. Generally speaking, they are found in arid and semi-arid lands. They are extensively found in such areas as the Great Basin of the United States, the Caspian region of Russia, and the interior of Australia. They come most prominently to the notice of soil investigators through their development in irrigated tracts. It is a common experience in such regions that soils deteriorate under irrigation. This may be due either to the rise of a saline water table or to the use of water containing excess of sodium salts. In many irrigated areas there are considerable tracts which have gone completely out of cultivation through alkalisation.

Whilst it is known that derelict alkali lands can be reclaimed by such methods as dressing with gypsum, in order to reconstitute the calcium soil, comparatively little has been done in practice to check deterioration in irrigated soils. The reclamation of such lands and the prevention of deterioration in land now under irrigated cultivation forms one of the most important tasks of applied pedology.

BIBLIOGRAPHY.

- ¹ MACKIE, W. W., "Reclamation of white-ash lands affected with alkali at Fresno, California." *U.S. Dept. Agric. Bur. of Soils*, Bull. 42, 1907.
- ² TUMIN, G.; from GLINKA, *Die Typen der Bodenbildung*, Berlin, 1914, p. 196.
- ³ GLINKA, K. D., *ibid.*
- ⁴ AARNIO, B., "Über Salzböden (Alaunböden) des humiden Klimas in Finnland." *Comp. Rend. IIIième Conference Agr.*, Prague, 1924, pp. 186-192.
- ⁵ VILENSKY, D. G., *Alkali Soils of Ukraina*, n.d., pp. 113-115.
- ⁶ SIGMOND, A. A. J. DE, "Contribution to the theory of the origin of alkali soils." *Soil Sci.*, 1926, **21**, pp. 455-475.
- ⁷ KOROTKY; in A. STEBUTT, *Lehrbuch der allgemeinen Bodenkunde*, Berlin, 1930, p. 404.

CHAPTER XIII.

SOILS OF THE HUMID TROPICS AND SUB-TROPICS*.

GENERAL CHARACTERISTICS.

BEFORE entering on a description of the principal groups of soils occurring in the humid tropics, it will be helpful to review the distinctive characteristics of soil formation under the conditions which obtain in these regions.

The most important circumstance differentiating the soils of the tropics from other soils is, of course, the temperature conditions under which they are developed. The mean temperature within the tropics is generally above 25°C , and in latitude 30° , is about 20°C . Bearing in mind that all chemical changes increase in velocity with rise of temperature, a rise of 10°C approximately corresponding with a doubling of velocity, it is easy to understand that the hydrolysis of mineral silicates, which forms the most important process in chemical weathering, proceeds with much greater rapidity under tropical conditions.

In addition, there has been an absence, during recent geological time, of glacial interference. Tropical weathering has thus operated over a much longer period of time than weathering in regions such as Northern Europe, where it can only date from the close of the last glacial epoch. There is also no interruption of chemical action by winter, as in temperate and cool regions. It is not surprising, therefore, that the depth to which epigene weathering extends is considerably greater than in temperate and cool climates. It is not uncommon to find fifty feet or more of weathered material.

*Other tropical soils are discussed in Chapters X, XI, and XIV.

Tropical soils may consist almost entirely of the products of chemical weathering, together with a greater or less admixture of unweatherable minerals such as quartz, magnetite, etc. Weatherable minerals such as feldspars and ferromagnesian minerals, are, in some cases, almost entirely absent. Such soils, if derived from basic rocks which do not contain quartz, may contain very high proportions of clay, as much as 80% being by no means uncommon.

With regard to rainfall, it is noteworthy that the equatorial regions are generally marked by high rainfalls and predominantly humid climates. Two peculiarities must, however, be noted.

Firstly, the rainfall tends to fall in heavy downpours, exceeding in intensity the heaviest rain encountered in higher latitudes. There are instances of as much as 6 inches of rain falling in an hour. This results in a marked tendency to erosion, particularly where the natural vegetative cover has been removed as a preliminary to cultivation. This leads to the development of extensive alluvial and colluvial fans in valleys. A further consequence of the prevalence of erosion is the production of truncated or immature profiles. Many of the difficulties which present themselves in the study of tropical soils might be obviated if the part played by erosion were more clearly appreciated.

Secondly, there is a general tendency for tropical humid climates to be divided into well marked wet and dry seasons. During the wet season, leaching by percolating waters may occur, resulting in a general impoverishment in bases and the production of acid profiles with consequent instability of the clay complex. The conditions during the dry seasons are totally different. By reason of the high temperature, evaporation is very intense and the soil becomes rapidly dried out to a considerable depth.

The prevailing colour of freely drained tropical soils is red, the actual tinge varying somewhat according to the amount of organic matter present and the mutual effect of the parent material and the pedogenic processes. In basin-

shaped areas with impeded drainage, and also where the parent material has a high lime status, dark coloured or grey soils may result. It may be assumed that both the processes affecting the inorganic weathering complex and the decomposition of plant residues proceed differently according as the soil profile is developed with free or restricted drainage.

The lime status also exerts an influence. Where the parent material is calcareous, the high concentration of calcium ions in the sphere of weathering prevents the loss of silicic acid, which otherwise occurs, and soils having a greyish mineral colour result. The clay fraction of such soils exhibits a relatively high silica-sesquioxide ratio.

Red soils appear, then, to result under conditions of free drainage and low lime status. Contrary to the impression produced by their bright red colour, such soils, until they are cultivated, may contain appreciable proportions of organic matter, particularly where the natural vegetation forms a close association such as forest, forest steppe, or savanna.

Soils subjected to submergence undergo characteristic changes, provided sufficient organic matter be present. Under these conditions, anaerobic decomposition occurs and rapid losses may ensue. There is also an increase in the solubility of the iron, manganese, calcium and magnesium, which appear in solution as bi-carbonates, the iron being in the ferrous state. The reductive processes result in a change from the red colour of well aerated soils to the grey colour associated with bottom soils. Where water movements can occur, the dissolved bases may be lost from the soil with consequent impoverishment.

The rôle of organic matter in tropical soil formation has been discussed by E. C. J. Mohr.¹ The amount of organic matter in a soil represents the balance between the addition of material in the form of plant residues and the destructive oxidation of added material by micro-organisms. The general relationship between plant growth and temperature may be expressed by a curve having its minimum at about

0°C, its optimum at 25°C and its maximum at something over 40°C. Given adequate moisture conditions, we may expect the maximum production of organic matter and, therefore, the maximum addition of plant residues at about 25°C. The curve connecting microbiological activity with temperature shows a different course. The minimum lies somewhat above that for the growth of the macro-flora. Mohr places it at about 10°C, but this is probably too high. The optimum lies at about 35-40°C, whilst the maximum may lie at 70-80°C, well above the temperatures encountered in actual soils. Mohr distinguishes microbiological decomposition under aerobic and anaerobic conditions, the intensity of decomposition being considerably greater under aerobic conditions.

Considering, now, the course of accession and destruction of organic matter in aerobic soils, the positive balance between the two processes increases up to about 20°C, after which the rate of destructive microbiological action increases more rapidly than plant growth with rise in temperature, until at somewhat above 30°C, destruction is sufficiently rapid to keep pace with additions of organic matter.

Under anaerobic conditions, less intense decomposition leads to the maximum balance between accession and destruction occurring at a higher temperature. The point at which gains are balanced by decomposition lies at about 35°C.

According to these ideas, humus should be present in very small proportions in virgin tropical soils with good aeration, and this may be broadly true for soils in which adequate moisture for biological decomposition is continually present. But it is not infrequently the case that virgin tropical soils contain organic matter in proportions comparable to those occurring in soils of temperate lands. Such soils are generally under steppe or savanna vegetation and it may be conjectured that the seasonal drought may have a conservative influence on soil organic matter as in the case of the tshernosems.

Except in regions of continuous rain, the humid tropics generally have marked wet and dry seasons. During the dry seasons, the soil moisture becomes relatively concentrated and tends to move upwards to the surface where it evaporates. There will thus occur a precipitation of materials from solution, principally in the surface horizon, which thus becomes an horizon of accumulation. The precipitation which occurs will be to some extent irreversible and result in the formation of the concretions which are such a distinctive feature of tropical soils. In the extreme case of laterite, the vertical transport and precipitation result in the development of a vesicular incrustation which forms the surface horizon. On the side of increasing humidity concretionary deposits will tend to become less frequent.

H. H. Bennett and R. V. Allison² have analysed the concretions (*perdigón*) from some typical Cuban soils. They are often markedly sesquioxidic in character but always contain a certain amount of silica. In some cases the silica may amount to over 50%. Such concretions are similar in composition to fairly siliceous clays.

RED LOAMS AND RED EARTHS.

These, the typical soils of the humid tropics, are developed under humid conditions with an alternation of wet and dry seasons. According to P. Vageler,³ the initial stage in the development of humid tropical soils, even where forest is later established, fulfils itself under steppe or savanna conditions. The high temperature ensures a rapid mineralisation of plant residues or else the formation of a type of humus of more neutral character than the humus of cooler climates. The reaction of the soil and, consequently, of the leaching is therefore neutral or only slightly acid. The instability of the weathering complex consequent on the removal of bases results in the liberation of silicic acid and sesquioxides. The neutral or slightly acid leaching leads to a removal of silicic acid, with the result that the residual material is relatively enriched in sesquioxides.

The extent to which the removal of silicic acid can proceed has not been clearly ascertained. It is true that in the case of laterites the soil may consist almost entirely of sesquioxides. But, in such cases, there is little doubt that the lateritic horizons are essentially of an illuvial character and do not represent residual material alone. From the writer's experience in Wales, it appears that, under temperate humid conditions, a weathering complex having a $\text{SiO}_2/\text{R}_2\text{O}_3$ of about 2.0 is produced by primary weathering and that any excess of sesquioxides over this ratio can be explained by illuvial accumulation. Analyses of the clays of tropical soils show $\text{SiO}_2/\text{R}_2\text{O}_3$ ratios considerably below 2.0; but whilst it is frequently asserted that a highly sesquioxidic residual—an allite in the terminology of Harrassowitz—is formed, it is by no means certain that the sesquioxidic soils so commonly encountered are not, like certain soils in Wales, partly illuvial in character.

If, however, it be assumed that weathering under humid tropical conditions can produce a residual complex with an excess of sesquioxides, variations may occur according to (a) the extent to which desilicification has occurred, and (b) the extent to which the soil or its individual horizons have received accessions of sesquioxides by illuviation. Apart from the most extreme laterites, which must be regarded as exceptional cases, tropical red soils always contain some silica in the clay complex and the presence of aluminosilicic clay constituents must be assumed.

The investigations hitherto made by X-ray methods on the clay complex have not disclosed any minerals having molecular $\text{SiO}_2/\text{R}_2\text{O}_3$ ratios less than 2.0. Whilst S. Mattson,⁴ by removal of excess sesquioxides, using hot aluminium chloride solution, obtained a residue of the halloysite type, the amounts of free sesquioxide found by F. Hardy⁵ in tropical soils would imply the presence of minerals having $\text{SiO}_2/\text{R}_2\text{O}_3$ ratios less than 2.0.

P. Vageler³ draws a distinction between *red loams* and *red earths*. We have seen that the tendency of weathering

under tropical conditions is towards the production of a weathering complex relatively enriched in sesquioxides by the removal of silicic acid or by accession of illuvial sesquioxides. In comparatively young soils these changes are not far advanced. The clay is of a siliceous type and the soil is described as a red loam. In the red earths, removal of silicic acid or accession of sesquioxides has proceeded sufficiently to give a weathering complex of a predominantly sesquioxidic character. A. Eichinger⁶ distinguishes three stages in soil development under humid tropical conditions, namely, younger red earth, older red earth, and laterite. All three types may occur under a uniform forest cover and the differences only become apparent after the institution of cultivation.

A red loam profile (Eichinger's younger red earth) consists of a deep and uniform layer of fairly plastic clay, often with yellow mottlings in the lower horizons and a gradual transition through a zone of decomposition to the parent rock.

Cuba. The soils of the Limones series in Cuba, described by H. H. Bennett and R. V. Allison⁷, may be cited as examples of the red loams. The profile consists of :—

- 0-4 in. Reddish-brown to dull red clay ; cracking into hard clods on drying.
- 4-12 (18) in. Pinkish-red clay, cracking in dry season. Fragments of parent rock present in lower parts of horizon.
- 12 (18)-24 in. Partly decomposed soft serpentine rock, greenish or bluish or hard rock.

The pH given is 6.72, the figure presumably referring to the surface soil. No data are available for the composition of the colloidal clay, but the silica and sesquioxide figures for the total soil of two successive layers are as follows :—

	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃
0-16 in.	51.25	17.40	10.69
16-42 in.	45.85	13.62	7.84

The clay is thus in all probability of a siliceous type, as might be expected from its plastic and cohesive character.

The relative shallowness of the profile and the absence of concretions suggest a soil in the early stages of development.

The Nipe Clay, on the other hand, is an example of a red earth, in which development has proceeded to an extreme stage. It is thus described by Bennett and Allison⁸ :—

- 0-26 in. Deep red friable clay with abundance of perdigón (concretions).
- 26-40 in. Red or brownish-red friable clay with less perdigón.
- 40 in.-13 ft. Yellow, exceedingly friable ochreous material.
- 13 ft.-16 ft. Reddish, yellowish, and almost black, exceedingly friable material extending to bed rock.
- 16 ft. + Serpentine, weathered in upper part, faced along fracture lines with thin coating of bluish to purplish black.

The pH varies about 6.0. The composition of the colloidal clay (G. Edginton) is :—

SiO ₂	Al ₂ O ₃	Fe ₂ O ₃
10.19	15.84	62.51

The highly sesquioxidic character of the clay is reflected in its friable character, whilst its maturity is indicated by the great depth of weathering. The exchangeable bases show the exceedingly low figures of 1.1 milligram equivalents per 100 g. of soil.

The Nipe clay is a highly weathered soil, but it is by no means certain that it represents simply a residual product. The abundant perdigón in the upper horizons certainly represents illuvial material, and it may be that the colloidal clay also is enriched by secondary accumulations of sesquioxides.

Central America. Friable clays of the red earth type are of frequent occurrence in Central America. H. H. Bennett⁹ describes a typical profile in the following terms :—

1. Surface layer 1/2 to 3/4 in. in thickness in which organic matter has slightly darkened the red soil.
2. Red clay uniform to 2-8 feet.
3. Red clay streaked or mottled with material of yellowish, bluish-grey, and whitish or cream colours often mixed with partly decayed rock material.

These soils are devoid of carbonates and low in exchangeable bases. The average silica-sesquioxide ratio for the total soil varies from 0.15 to 1.96.

E. Africa. The writer has been fortunate, through the courtesy of Mr. G. Milne, of the Amani Institute, Tanganyika, in securing a number of typical red soils from East Africa. The analytical data obtained from them present many points of interest.

The first profile to be discussed is from Muhesa, Tanganyika. It occurs on the breast of a long slope at about 600 ft. altitude, under a rainfall of about 50 inches per annum. Mr. Milne's description of this profile, which is that of a typical red soil, is as follows:—

“Reddish chocolate top soil, 0-5 inches; red clay below; quartz gravel layer at 30-36 inches; clay again below that, becoming yellow-brown; then yellow to rock-brash and then raw gneiss.”

The analytical data are shown in Table XXXIII.

TABLE XXXIII.—ANALYTICAL DATA FOR MUHESA SOIL.

Horizon	Clay %	Organic Carbon	pH	Clay Fraction				
				SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
0-15 in.	55.25	1.65	7.7	40.5	30.6	17.6	2.25	1.64
12-18 in.	70.94	0.31	6.6	44.4	33.9	17.8	2.23	1.89

A profile from Kwamshindi, Tanganyika, occurring at 2,700 feet with an annual rainfall of 80 inches, under secondary bush, formerly rain forest, is thus described:—

“Dark coloured chocolate subsoil, cubical structure, much angular quartz gravel; passes through orange and drab-yellow (2 feet thick) to raw gneiss at about 12 feet.”

The analytical data are shown in Table XXXIV.

TABLE XXXIV.—ANALYTICAL DATA FOR KWAMSHINDI SOIL.

Horizon	Clay %	Organic Carbon	pH	Clay Fraction				
				SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
0-12 in.	45.2	3.21	6.3	44.2	37.5	15.8	2.01	1.58
12-20 in.	45.05	1.89	6.0	42.8	37.3	16.4	1.95	1.53

Another profile from gneiss near Amani, at 3,300 feet, rainfall 85 inches, in virgin forest is described as

“ Dark blackish-brown top soil, 0.8 inches: dark orange-brown subsoil, much rock-brash, at 27 inches and below; raw gneiss probably at about 6.8 feet.”

The analytical data are shown in Table XXXV.

TABLE XXXV.—ANALYTICAL DATA FOR AMANI SOIL.

Horizon	Clay %	Organic Carbon	pH	Clay Fraction				
				SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
0.3 in.	27.4	6.17	6.1	26.3	33.5	32.4	1.33	0.8
12-18 in.	40.6	1.72	5.1	24.6	31.4	34.8	1.33	0.78

An important fact which emerges from these analyses and those of a large number of other soils from East Africa is that larger proportions of organic matter, as judged from the organic carbon, are present than would be thought possible under tropical conditions. The soil from virgin forest at Amani contains over 6% organic carbon, corresponding with about 10-11% of organic matter. Even under secondary bush, the Kwamshindi soil contains the respectable content of 3.21% organic carbon.

The figures for the composition of the clay fraction in the first two soils do not indicate a pronounced lateritic development. They might be paralleled by many figures for moderately eroded soils in the uplands of Wales. The third example, if it can be assumed that no severe erosion has taken place, indicates a definitely lateritic type of weathering complex. But even here the sesquioxidic character is mainly due to the high proportions of ferric oxide.

S.E. United States. Red soils are common throughout the South-Eastern United States and may have affinities with the red soils of the tropics.

A profile near Greensboro, N. Carolina, described by R. C. Journey¹⁰, showed the following characteristics:—

- A. 0 to 9 in. Slightly reddish-brown clay loam.
- B₁. 9 to 36 in. Deep red heavy brittle clay.
- B₂. 36 to 60 in. Light red friable crumbly clay.
- C. 60 in. + Ochreous yellow, black and reddish-brown decomposed diorite.

There is a marked accumulation of clay in the B₁ horizon. The organic matter decreases from the surface, and the pH falls from 6.3 in the A horizon, to 4.1 in the C horizon. The silica-sesquioxide ratios in the colloidal clay of the successive horizons are 1.46, 1.49, 1.42, and 1.40, respectively. Ferric oxide increases with depth, the figures being 15.71, 18.94, 24.13, and 23.64%, respectively.

L. D. Bayer and G. D. Scarseth¹¹ have discussed the genesis of red soils in Alabama, with special reference to soils of the Tusquehanna series. They consider that the northern limit of the type of weathering characterised by sesquioxide enrichment corresponds with the 61°F mean annual isotherm.

Southern China. Red soils in Southern China are described by C. F. Shaw¹². They occur under a mean annual rainfall of 50-60 inches. A typical profile under small herbs and grass consisted of 2-5 inches of brownish red clay loam, friable but with very little organic matter, over a clay subsoil of a red colour with reticulate yellow and grey mottling. The parent material is shale and sandstone.

Indo-China. M. V. Agafonoff¹³ reports red soils from Indochina. They are derived from basaltic rocks, and the ferruginous crusts which are developed in certain cases are considered to result from upward translocation of sesquioxides during drought.

FRIABLE AND NON-FRIABLE SOILS.

It is possible to draw a distinction between friable and non-friable tropical soils. The former appear to belong to Vageler's red earths and Eichinger's older red earths and laterites, and the latter to Vageler's red loams and Eichinger's younger red earths. They are distinguished

principally by the character of their weathering complex, the non-friable soils being relatively siliceous, whilst the friable soils are relatively sesquioxidic in character.

Siliceous and sesquioxidic soils are strongly contrasted in their textural properties. In the former type, the weathering complex is definitely of a clay character and, where present in considerable amounts, gives rise to the characteristic plasticity and cohesiveness of clay soils. In soils whose clay complex contains excess of sesquioxides, the properties of plasticity and cohesiveness are less strongly marked. A soil may contain very high proportions of this weathering complex and yet be of an open friable character. This distinction has been brought out strongly by the work of H. H. Bennett⁹ in Central America.

Comparing friable and non-friable soils, Bennett obtained the following data (Table XXXVI).

TABLE XXXVI. — AVERAGE AND RANGE OF SILICA, FERRIC OXIDE, ALUMINA, TOTAL BASES, AND $\text{SiO}_2/\text{R}_2\text{O}_3$ IN FRIABLE AND NON-FRIABLE SOILS (H. H. BENNETT).

	SiO_2	R_2O_3	$\text{RO} + \text{R}_2\text{O}$	$\text{SiO}_2/\text{R}_2\text{O}_3$
Friable ...	29.30(5.00-41.87)	49.19(31.19-66.72)	1.32(0.13-6.55)	1.25(0.15-1.96)
Non-friable	52.33(41.27-77.00)	30.81(12.18-38.78)	4.50(0.85-17.17)	3.71(2.00-11.60)

The soils of the humid tropics and, indeed, of the humid sub-tropics, appear to present all stages between laterite and soils in which free sesquioxides, even if present, do not dominate the properties of the soil, although their presence may be inferred from the red colours which generally obtain.

AFFINITIES OF TROPICAL RED SOILS WITH BROWN EARTHS.

The tropical red soils hitherto described, with the possible exception of the red soils from Southern China, are derived by primary weathering from crystalline rocks. Much of the material on which current ideas of soil classification are based consists of studies on soils of secondary

weathering, i.e., in which products of chemical weathering are already present in the parent material. It is therefore important to bear in mind the possibility that some of the characteristic features of the soils of the humid tropics might be found in soils of primary weathering in temperate climates. Apart from the greater degree of weathering and the occurrence of concretionary material, the writer sees no essential difference between the reddish-brown soils derived from crystalline rocks in Britain and the younger red loams of East Africa. It must be remembered that, even in the humid tropics, concretionary material is not invariably present and, except in ground-water soils, only becomes an important constituent under climatic conditions marked by intensive drought.

The more vivid red colours of tropical soils may be attributed to difference in the degree of hydration of the ferric oxide. Whilst, under temperate climates, the hydrates are of the limonitic type, the less hydrated turgite, goethite, and even hæmatite, are probably present in tropical soils.

YELLOW EARTHS.

This is a rather ill-defined group of soils of which representatives may occur in the tropics but which is more commonly encountered in sub-tropical and warm temperate climates. The yellow colour is probably connected with a higher degree of hydration of the ferric oxide in these than in the red soils. Yellow earths probably form a transition group between the tropical red soils and the brown forest soils of temperate climates. Certain yellow soils, such as the Norfolk, in the S.E. United States, are considered by C. F. Marbut to differ from adjacent red soils on account of the removal of iron during an earlier stage of impeded or sluggish drainage.

Yellow soils may also result from incipient podsolisation of red soils or may represent early stages in the development of red soils. For the present it would be unwise to attempt to separate them into a definite soil group.

LATERITE.

In 1807, H. Buchanan¹⁴ described a type of red soil occurring frequently in Southern India and proposed for it the name *laterite* (L. *later*=brick). The term has been widely applied to tropical red soils of widely differing characters and no problem in soil genesis has been more earnestly debated than the nature and mode of origin of this material.

One of the principal difficulties under which this branch of pedology has laboured is the fact that many of the descriptions of the supposed laterites merely relate to red soils. Indeed, the term laterite is often used synonymously with red soil. Red soils are certainly characteristic of the humid tropics, but to class them all as laterites would be to assemble in one group soils which would be distinct in any natural system of classification. It is now generally agreed that the terms laterite and lateritic should be restricted to materials characterised by excess of sesquioxides.*

Some writers would define laterite in terms of alumina. For example, F. J. Martin and H. C. Doyne¹⁵ would define laterite as material in which the molecular ratio of silica to alumina in the clay is less than 1.33, whilst in lateritic soils the ratio is 1.33 to 2.0.

A typical laterite profile appears to consist of four horizons, namely: (1) the parent rock succeeded by (2) an horizon in which the material is non-lateritic and the immediate product of primary weathering; (3) the horizon of laterite proper, which passes into (4) an horizon characterised by ferruginous incrustations or concretions. The zone of decomposition may attain a thickness of a metre or more, or may be a mere film separating the rock from the lateritic zone.

The lateritic horizon consists of red material, often mottled with yellow and even violet, generally argillaceous, but, when derived from quartzose rocks, often sandy in

*For a discussion of the use of the term laterite, see L. L. Fermor (Geol. Mag., 1911, V, 8, 454-462, 507-516, 559-566).

texture. It has a tendency to a vesicular or cellular structure and the pores are frequently filled with white or greyish material. When moist it may be dug with a spade, but on drying it sets to hard lumps which are sometimes used as a building material.

The surface horizon of concretionary material may attain such a development as to preclude the growth of vegetation. Below the superficial crust there is generally a certain separation of hydrated alumina (hydrargillite or gibbsite) which attains its greatest development in the bauxite-laterites.

F. Hardy and R. R. Follett-Smith¹⁶ describe a number of lateritic profiles from British Guiana. The following profile, derived from hornblende schist, is typical:—

1. Superficial laterite and ironstone, frequently vesicular and slaggy, scattered in blocks on the surface.
2. Topsoil. Red-brown humic clay, 9 inches.
3. Tawny red earth, friable and very gravelly, especially in its upper part. The gravel consists of pieces of laterite, ironstone pellets, and quartz fragments. Thickness 10-20 feet.
4. Primary red laterite, hard and crusty. Thickness 3 inches over broken rock surface.

Gibbsite ($\text{Al}_2\text{O}_3 \cdot 3\text{H}_2\text{O}$) occurs throughout the profile. It would appear that in this type of profile laterite is the immediate product of primary weathering. F. J. Martin and H. C. Doyne¹⁷ have also recorded the direct formation of lateritic material from norite in Sierra Leone.

It is considered by J. Harrison (unpublished M.S. quoted by Hardy and Follett-Smith) that two main processes may be distinguished, namely, (1) primary lateritic formation and (2) resiliification of the lateritic product by deposition of silica from ascending solutions. The latter process only takes place in localities affected by a high water table. Otherwise, the laterite appears to be permanent.

Hardy and Follett-Smith adduce evidence for a process similar to podsolisation affecting the upper horizons of lateritic profiles, whereby sesquioxides are removed by leaching.

J. M. Campbell¹⁸, from a study of Australian and African laterites, considers laterites to result from changes which occur in the zones of permanent and intermittent saturation in the vicinity of a water table. Since oxygen is necessary for the oxidation of ferrous to ferric iron, the proximity of these zones to the surface is a further condition for its formation.

The hydrated sesquioxides deposited in the lateritic horizons originate by hydrolysis of silicate minerals and occur in the ground-water in alkaline solution, the iron as ferrous hydrogen-carbonate and the aluminium as alkaline aluminates. During periods of drought, there is an upward translocation of dissolved material from the water table. Ferrous hydrogen-carbonate undergoes oxidation, giving amorphous hydrated ferric oxide, whilst the aluminates give rise to amorphous hydrargillite $\text{Al}_2\text{O}_3 \cdot 3\text{H}_2\text{O}$.

Subsequent changes in the lateritic horizon involve the formation of turgite $2\text{Fe}_2\text{O}_3 \cdot \text{H}_2\text{O}$ and possibly hæmatite, Fe_2O_3 , from amorphous ferric oxide hydrate, and crystalline gibbsite $\text{Al}_2\text{O}_3 \cdot 3\text{H}_2\text{O}$ from hydrargillite. On account of the greater mobility of ferric oxide in colloidal solution, it may happen that lateritic horizons become progressively more aluminous in character with age.

The views of Campbell are supported by the investigations of W. G. Woolnough¹⁹ in Western Australia. This author considers that laterisation can only occur in situations in which free drainage is at a standstill, as in a peneplain at or about sea level. High level laterites are products of processes which have long since ceased consequent on uplift and the alteration of hydrological conditions. Laterite, as imagined by these authors, was formed by a process which may be paralleled by the formation of "murrum" and "mocarrero" in contemporary vleis soils.

Such a view may be contrasted with Harrison's theory of primary laterite formation. It is inadequate to explain the apparently direct laterisation of norite rock in Sierra Leone described by F. J. Martin and H. C. Doyne¹⁷, and is

obviously inapplicable to the development of the ferruginous Nipe clay of Cuba.

The formation of a material such as the Nipe clay, which appears to be mainly residual in character, must obviously be distinguished from the processes whereby profiles characterised by slaggy ferruginous incrustations are developed. In these, the more typical lateritic profiles, it is clear that there has been a deposition which has masked the primary processes of silicate hydrolysis and clay formation. It cannot be assumed in all cases, however, that profile development has been influenced by the proximity of ground-water,* for contemporary laterites may occur under savanna.

Whether illuvial deposition of sesquioxidic material has taken place in the presence or absence of a high ground-water table, it is evident that the principal climatic circumstances are high temperature and the alternation of wet seasons with intense drought.

The mobility of sesquioxidic constituents in the soil profile has been attributed by Campbell to the alkalinity of the ground-water, by H. Harrassowitz²⁰ to the protective effect of humus sols and by A. Reifenberg²¹ to the protective effect of silicic acid sols. It is probable that each of the three explanations may find an application in the wide range of lateritic profiles encountered.

The parent material appears to have considerable significance, for, according to Hardy and Follett-Smith, acidic rocks such as granite do not yield laterites. This may be connected with the formation of a kaolinitic type of weathering product from the potash and potash-soda feldspars occurring in such rocks.

*The name lake laterite is suggested by L. L. Fermor (Geol. Mag., 1911, V, 8, 454-462) for lateritic depositions in the vicinity of a water table or free water surface.

BIBLIOGRAPHY.

- ¹ MOHR, E. C. J., *Tropical soil forming processes and the development of tropical soils*, trans. by R. C. PENDLETON, Coll. Agric. Univ. of Philippines, 1930.
- ² BENNETT, H. H., and ALLISON, R. V., *The soils of Cuba*, Washington, 1928, p. 79.
- ³ VAGELER, P., *Grundriss der Tropischen und Sub-tropischen Bodenkunde*, Berlin, 1930.
- ⁴ MATTSON, S., "Laws of colloidal behavior." V.—Ion absorption and exchange. *Soil Sci.*, 1931, **31**, pp. 311-331.
- ⁵ HARDY, F., "Studies in tropical soils." I.—Identification and approximate estimation of sesquioxide components by adsorption alizamine. *J. Agric. Sci.*, 1931, **21**, pp. 150-166.
- ⁶ EICHINGER, A., "Die Entstehung der Roterde und Laterite." *Z. Pflanz. Düng.*, 1927, **8A**, pp. 1-13.
- ⁷ BENNETT, H. H., and ALLISON, R. V., *Soils of Cuba*, p. 144.
- ⁸ BENNETT, H. H., and ALLISON, R. V., *Soils of Cuba*, p. 238.
- ⁹ BENNETT, H. H., "Some comparisons of the properties of humid-tropical and humid-temperate American soils; with special reference to indicated relations between chemical composition and physical properties." *Soil Sci.*, 1926, **21**, pp. 349-374.
- ¹⁰ JURNEY, R. C., in M. S. ANDERSON and H. G. BYERS, *U.S. Dept. Agric. Techn.*, Bull. 228, 1931.
- ¹¹ BAVER, L. D., and SCARSETH, G. D., "Subtropical weathering in Alabama as evidenced in the Sesquehanna fine sandy loam profile." *Soil Res.*, 1931, **2**, pp. 288-307.
- ¹² SHAW, C. F., "Soils of China" *Geol. Survey of China*, 1930, Bull. 1, pp. 1-38.
- ¹³ AGAFONOFF, M. V., "Sur quelques sols rouges et Bienhoa (sols) de l'Indochine." *Soil Res.*, 1930, **2**, pp. 184-196.
- ¹⁴ BUCHANAN, H., *Journey from Madras through Mysore, Canara, and Malabar*, 1807.
- ¹⁵ MARTIN, F. J., and DOYNE, H. C., "Laterite and lateritic soils in Sierra Leone." *J. Agric. Sci.*, 1927, **17**, 530-547.
- ¹⁶ HARDY, F., and FOLLETT-SMITH, R. R., *J. Agric. Sci.*, 1931, **21**, pp. 739-761.
- ¹⁷ MARTIN, F. J., and DOYNE, H. C., "Laterite and lateritic soils in Sierra Leone," II. *J. Agric. Sci.*, 1930, **20**, pp. 135-143.
- ¹⁸ CAMPBELL, J. M., "Laterite, its origin, structure, and minerals." *Mining Mag.*, 1917, **17**, pp. 67-77, 120-128, 171-179, 220-229.
- ¹⁹ WOOLNOUGH, W. G., "Laterite in Western Australia." *Geol. Mag.*, 1918, VI, **5**, pp. 385-393.
- ²⁰ HARRASSOWITZ, H., "Laterit." *Fortschr. Geol. Palaeont.*, 1926, **4**, pp. 253-566.
- ²¹ REIFENBERG, A., "Über die Rolle der Kieselsäure als Schutzkolloide bei der Entstehung mediterraner Roterde." *Z. Pflanz. Düng.*, 1927, **10A**, pp. 159-186.

CHAPTER XIV.

SOILS ASSOCIATED WITH CALCAREOUS PARENT MATERIALS.

THE soils to be discussed in the present chapter are grouped together principally for descriptive purposes and not, necessarily, because they form a major group in a world system. Their position in such a system will be discussed at a later stage.

Soils derived from or associated with calcareous parent materials, such as limestone, chalk, or coral, appear to fall into two main groups. On the one hand, is a group of soils in which the mineral colour is grey or greyish-brown, and the colour of the soil as a whole tends towards brownish or greyish-black shades. With them may be grouped white or grey chalk soils in which the organic matter colour is masked by the excess of chalk. Soils of this group occur in temperate, sub-tropical, and tropical climates. On the other hand, is a group of soils in which the colour is predominantly red or reddish-brown, sometimes modified by the presence of organic matter, which would appear to be of a different character from the organic matter of the soils of the first group. The clay complex of the black and grey soils is more siliceous than that of the red and brown soils, which appears to contain free sesquioxides. Soils of the first group contain free calcium carbonate and are completely saturated with exchangeable bases, whilst soils of the second group show a distinctly lower base status and may be actually acid in reaction. To the first group belong the rendzinas, with which may be included the grey and white limestone soils of the temperate regions, and the black lime-

stone and marl soils of the tropics. To the second group belong the so-called terra rossa soils, with which may be included the brown and red limestone soils of temperate regions and the red limestone soils of the tropics.

RENDZINAS AND RELATED SOILS.

These soils have been most widely studied in Central Europe. They are sometimes described as humus-carbonate soils. They are, typically, dark coloured soils containing from 3-12% of organic matter and varying amounts of free calcium carbonate. In Glinka's system of classification, they are considered to be typical endodynamomorphic soils, which have not yet reached mature development and in which the profile character is mainly determined by the nature of the parent material. Though generally of a dark colour, soils belonging essentially to this class may also be dull brown or even grey in colour. Reddish-brown colours do not occur.

According to K. D. Glinka¹ the essential features of a rendzina profile are :—

- A₁. Grey, dark grey, or almost black soil, sometimes containing fragments of chalk or marl. Thickness 15-30 cm.
- A₂. Whitish-grey, slightly humus coloured or brownish, with more fragmental chalk or marl than in the A₁ horizon.
- C. Calcareous parent rock.

Poland. S. Miklaszewski² distinguishes two groups of rendzinas, namely, those derived from calcareous parent materials and those from gypseous parent materials. Among the calcareous rendzinas, white, brown, and black varieties occur. The colour is not always correlated with humus content, for an instance is given of a white rendzina, containing 3.5% of humus, adjacent to a black rendzina with 2.8% of humus. The white rendzinas are probably similar to the white chalk soils of England, in which the organic matter colour is masked by the excess of chalk. The presence of considerable proportions of calcium carbonate is a constant feature of typical rendzinas.

England. The writer is occupying a debatable position in assigning the chalk soils of England to the rendzina

group, but the prevalence of grey or brownish-grey colours and the presence of free calcium carbonate would appear to warrant their inclusion. Even in the light coloured chalk soils, the residue after removal of calcium carbonate has the dull greyish-black colour characteristic of organic matter formed in the presence of excess of calcium carbonate.

The chalk soils of Great Britain generally appear to have undergone considerable erosion, and this has tended to obscure their relationship to the continental rendzinas. Yet, under certain conditions, profiles occur in which there is a well marked greyish-brown or even black surface horizon.

A chalk soil profile, occurring under pasture near Wantage, Berkshire, is described by Miss F. F. Kay³ in the following terms:—

- 0-10 in. Grey fine sandy silt loam.
- 10-16 in. Putty-coloured fine sandy silt loam containing many fragments of chalk rock, which increase in size with depth.
- >16 in. Chalk rock.

The following analytical data were obtained on samples of the above profile (Table XXXVI).

TABLE XXXVI.—ANALYTICAL DATA FOR CHALK SOIL PROFILE.

Horizon	CaCO ₃ %	Organic Carbon %	Clay Fraction				
			SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
0-10 in.	19.2	2.3	57.0	15.0	7.8	6.46	4.82
10-16 in.	45.2	0.7	52.9	15.8	7.0	5.69	4.46

Wales. Certain soils derived from the Lower Lias would appear to belong to the rendzina group. The following is a description by D. O. Hughes⁴ of a typical profile in Glamorganshire under old pasture:—

- 0 to 6 in. Greyish-brown stoneless clay loam.
- 6 to 15 in. Light yellowish-brown clay, with occasional fragments of limestone.
- >15 in. Soft limestone.

The following analytical data were obtained (Table XXXVII).

TABLE XXXVII.—ANALYTICAL DATA FOR LIAS SOIL PROFILE.

Horizon	Clay %	CaCO ₃ %	Organic Carbon %	Clay Fraction				
				SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
0-6 in.	39.5	2.7	3.3	55.9	21.1	8.8	4.50	3.56
6-15 in.	44.0	17.7	1.2	55.3	21.9	8.0	4.29	3.47

The highly siliceous character of the clay fraction is a distinctive feature of soils belonging to this group which have been examined by the writer. On theoretical grounds, the same character may be expected in all soils containing an excess of calcium carbonate, maintaining a state of base-saturation.

Cuba. The Camaguey series of Cuba described by H. H. Bennett and R. V. Allison⁵ may serve as an example of a tropical black limestone soil. Typically it consists of (1) 6-14 in. of dark brown to black clay, overlying (2) dark brown to yellowish-brown plastic clay, frequently calcareous, grading at 20-30 in. into (3) brownish-yellow to yellow, plastic and somewhat sticky clay, frequently calcareous and containing lumps of soft calcium carbonate, overlying at 40-72 in. (4) either whitish cocó (chalky limestone), pale yellow calcareous clay and white cocó, or cocó and calcareous clay, interbedded with hard or semi-hard limestone.

The analytical figures given show that calcium carbonate is present throughout the profile. The organic matter in the surface soil is 9.62%.

RED AND BROWN LIMESTONE SOILS.

Terra rossa is the name given to a red soil which occurs commonly in the countries bordering on the Mediterranean Sea. Typically it is associated with limestone and in the present account we shall follow the definition of A. Reifenberg⁶ in restricting the term terra rossa to such soils.

The soils included in this group show a simple type of profile consisting of a variable depth of red soil sometimes slightly modified in colour by admixture of humus in the upper horizon, and passing abruptly to the limestone rock below without any marked zone of transition. Typically, the terra rossa soils are fairly heavy clays. Analyses by W. Graf zu Leiningen⁷ show clay contents varying from 32.85 to 59.05% in typical examples. The coarser fractions are correspondingly small. In some cases, fragments of the parent rock and also calcareous concretions stained red with iron oxides are present.

Although numerous bulk analyses are found in the literature, scarcely any data are available for the composition of the clay fraction. A clay from Pikermi examined by zu Leiningen gave a clay fraction having the composition,

SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	CaO	Ignition
41.62	21.16	12.85	3.82	15.05

Similar figures were obtained by E. Blanck, F. Kunz, and F. Preiss⁸ for certain red soils from Moravia with resemblances to terra rossa. The present writer has made partial analyses of the clay fractions of two red soils over limestone from Cap d'Antibes and St. Gowans Head, Pembrokeshire, respectively. The figures are as follows:—

	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃
Cap d'Antibes ...	47.9	31.2	15.4
St. Gowans Head	44.9	30.4	13.6

Neither the abundant data for the bulk composition of terra rossa soils nor the few available data for the composition of the clay fraction suggest any tendency towards laterisation. The figures are such as might be obtained from typical brown earths, or younger red tropical loams.

Generally speaking, terra rossa soils are base-saturated. Calcium carbonate is, indeed, shown in certain analyses, but it is not improbable that it may be present in the form of fragments which, by reason of their size or degree of induration, are without effect on the base status of the body of the soil. In this connexion, it has been re-

marked by K. D. Glinka, that whilst soft limestones produce rendzina soils, hard limestones produce red soils. If the development of red soils be a consequence of a lowering of base status, it is readily seen that such a process can most readily take place in soils derived from hard limestones.

The problem of the origin of terra rossa has been earnestly debated both by pedologists and geologists. The theory that it is simply a residual product from the parent limestone is inadequate, since material resembling terra rossa cannot be obtained by dissolving away the carbonates from limestones. And, therefore, whilst its close association with the limestone suggests an origin from the non-carbonate portion of the rock, it is evidently necessary to account for the changes whereby this material has become terra rossa.

H. Stremme⁹ considers terra rossa to be analogous to the B horizon of podsol profiles, the A horizon having been removed by erosion. But whilst red subsoils similar to terra rossa occur in certain parts of Central Europe under humus surface soils, it does not appear possible to apply this theory of their origin in the Mediterranean region generally.

A. Reifenberg⁶, from an intensive study of the conditions of terra rossa formation, concludes that the essential feature in its formation is the irreversible precipitation, in the surface horizons, of sesquioxide sols peptised by silicic acid. Such data as are available, however, do not suggest a high content of sesquioxides for terra rossa, and it may be more correct to regard these soils as analogous to the brown earths, in which, if we accept the views of A. Stebutt, the lowering of the base status has resulted in a partial degradation of the zeolitic or clay complex into its component silicic acid and sesquioxides.

For a fuller understanding of the process of terra rossa formation it is highly desirable that data should be obtained for the composition of the clay fraction in typical profiles.

Cuba. As an example of a tropical red limestone soil, the Matanzas series of Cuba described by H. H. Bennett and R. V. Allison¹⁰ may be cited.

The profile is comparatively simple and consists of a uniform red friable clay, which may reach to 15 feet or more with an abrupt transition to the parent rock. The top 3-6 inches is somewhat darker in colour through the presence of organic matter. A characteristic of this series is the occurrence of concretionary material (perdigón). The parent rock is light coloured limestone, locally flinty, or limestone and cocó.

The analytical figures presented show a complete leaching of calcium carbonate, a condition which is reflected in the acid reaction of the soils of this series. The organic matter content in the three surface soils for which data are given varies from 2.19 to 4.65%. Although no data for the composition of the clay fraction are available, free sesquioxides are probably present. The concretions consist mainly of ferric oxide with smaller proportions of alumina and silica.

United States. The Hagerstown Series which is found in the Blue Ridge region of the United States belongs to the group of red and brown limestone soils. The soil profile, where mature, consists of a light-textured light brown A horizon and a brown or reddish-brown B horizon. Calcium carbonate has been entirely removed from the profile by leaching.

Wales. Brown, red, and reddish-brown soils are commonly found on hard Carboniferous Limestone in Wales under conditions which have resulted in decalcification and desaturation. It is noteworthy that, under the same climate, the soft Lias Limestone gives rise to greyish-brown soils. The following profile is found on Carboniferous Limestone at Bridgend, Glamorgan :—

- 0-9 in. Rusty reddish-brown silty heavy loam.
- 9-15 in. Brown silty heavy loam with fragments of limestone.
- >15 in. Hard limestone.

Fragments of limestone occur in both horizons and markedly in the second horizon. Yet the soil is partly leached, for the pH is 5.96 and the exchangeable lime

figures, 0.379% and 0.313% respectively for the two horizons, indicate partial desaturation. The composition of the clay fractions is as follows:—

	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	SiO ₂ /Al ₂ O ₃	SiO ₂ /R ₂ O ₃
0-9 in.	47.5	32.2	10.7	2.50	2.07
9-15 in.	48.2	29.6	12.6	2.76	2.17

These figures show a marked contrast with those for neighbouring greyish-brown Lias soils, which appear to belong to the rendzina group.

Whilst in ordinary Welsh profiles, the tendency is for sesquioxides to increase in the clay fraction from the surface downwards, there is here no evidence of eluviation in either direction.

The fairly high proportions of organic matter in the desaturated limestone soils of Wales tend to mask their mineral colour, which is often decidedly brick red. More commonly the colour is a warm reddish-brown.

RELATIONSHIP OF RENDZINAS TO RED AND BROWN LIMESTONE SOILS.

We have seen that limestones can give rise to two distinct groups of soils. The one group is characterised by a high base status, excess of calcium carbonate, a dark coloured type of organic matter, and a grey or greyish-brown mineral colour associated with a siliceous type of clay, probably devoid of free sesquioxides.

The other group is characterised by a low base status, sometimes even in the presence of fragmentary limestone, a type of organic matter of a lighter colour than that of the first group, and a red or reddish-brown mineral colour, associated with a type of clay having a SiO₂/R₂O₃ ratio of 2.0 or less, in which there is a certain proportion of free sesquioxides.

The formation of the second group appears to follow the desaturation of the soil, and we may therefore regard the red limestone soils as more mature than the grey or rendzina soils. In a limestone country, it should be possible to dis-

cern different stages of maturity due to variations in topography. If mature leached soils are to be found, they may be expected in upland situations with relatively high rainfall and intense leaching, provided erosion is not severe enough to keep pace with profile development.

An example of this is furnished by red and black soils derived from coral limestone in Barbados, described by S. J. Saint¹¹. The red soils occur in the districts of highest altitude and greatest rainfall, whilst the black soils occur in the drier lowlands. All intermediate gradations are found between the extremes. The black soils are more colloidal than the red soils, and preliminary data show that they have a more siliceous clay fraction. It is rather remarkable that the red soils have a higher organic matter content than the black soils. The dark colour of the latter soils is probably associated with the higher calcium status of the humus, a circumstance which has been remarked elsewhere. According to Saint, the key to the distinction between the two types lies in the greater maturity of development of the red soils.

In the writer's experience, soils derived from hard limestones in Wales under conditions of intense leaching always show reddish or brownish colours, whilst colluvial soils or soils in situations with imperfect drainage are generally greyish or greyish-brown.

In view of these considerations, it may be permissible to hazard a conjecture that the red Clay-with-flints overlying certain parts of the Chalk of S.E. England may represent a mature limestone soil. Against the view that it is a residual product from chalk it has been urged that the chalk which has been weathered away could not have produced a layer of Clay-with-flints of the thickness which actually occurs. Yet a considerable proportion of the clay must be chalk residual material, and it is not impossible that some additions may have been made by illuvial accumulation as has been suggested for the terra rossa.

It may be remarked that mature limestone soils tend to resemble the soils of warmer climates. The Mediterranean

terra rossa shows points of resemblance with the red loams of the tropics, whilst the red colours of leached limestone soils of more northern climates are in marked contrast to the more sober brown colour of adjacent soils derived from non-calcareous parent materials. This may be attributed to the drier and, therefore, warmer character of these soils and to the more rapid mineralisation of plant residues.

BIBLIOGRAPHY.

- ¹ GLINKA, K. D., *Die Typen der Bodenbildung*, Berlin, 1914, p. 215.
- ² MIKLASZEWSKI, S., "Contribution à la connaissance des sols nommés 'rendzinas.'" *Compt. Rend. IVième Conférence Agro-pédologique*, Prague, 1924.
- ³ KAY, F. F., Private communication.
- ⁴ HUGHES, D. O., Private communication.
- ⁵ BENNETT, H. H., and ALLISON, R. V., *The soils of Cuba*, Washington, 1928, p. 49.
- ⁶ REIFENBERG, A., "Die Entstehung der Mediterran-Roterde." *Kolloid-chem. Beihefte*, 1929, **28**, pp. 1-93.
- ⁷ ZU LEININGEN, W. GRAF, "Entstehung und Eigenschaften der Roterde." *Int. Mitt. Bodenk.*, 1917, **7**, pp. 176-204.
- ⁸ BLANCK, E., KUNZ, F., and PREISS, F., "Über mährische Roterde." *Landw. Vers. Stat.*, 1923, **101**, pp. 247-260.
- ⁹ STREMMER, H., "Laterit und Terra rossa als illuviale Horizonte humoser Waldböden." *Geol. Rdsch.*, 1914, **5**, p. 480, *et seq.*
- ¹⁰ BENNETT, H. H., and ALLISON, R. V., *The soils of Cuba*, Washington, pp. 26-28.
- ¹¹ SAINT, S. J., *Rept. Dept. Sci. Agric. Barbados*, 1930-31, pp. 70-73.

CHAPTER XV.

THE CLASSIFICATION OF SOILS.

IN considering the problem of soil classification, it should be realised at the outset that there are many possible systems, each of which may have a value for a particular limited purpose. From the philosophical standpoint, however, no system of classification can be considered satisfactory which fails to express the genetic relationships of soils to each other. For example, a classification of soils according to texture, which would be quite adequate for practical purposes in a limited area, might lead to soils being grouped together which are completely unrelated genetically. The earlier attempts at soil classification were generally made with a view to their practical significance, and it is only during comparatively recent years that the purely scientific viewpoint has secured general recognition in Western Europe and America.

Perhaps the chief obstacle in arriving at a system of soil classification is the inevitably limited experience of individual workers. Thus, systems of classification which are philosophically valid for limited regions may prove to be inapplicable for world-wide use. F. A. Fallou¹, from his experience in Saxony, devised a classification of soils based on the character of the parent material. But such a system is inapplicable to the world classification because it neglects the effect of climate on the processes of soil formation. The same rock may give rise to widely different soils when subjected to a variety of pedogenic processes.

Another defect in earlier systems of soil classification lay in the restriction of attention to the surface soil. It is now realised, as the consequence of the work of the Russian school of pedologists, that the complete expression of the

pedogenic processes, acting on a given parent material, is found in the soil profile, consisting of all the horizons differentiated by these processes from the substrate. The unit of study and classification is therefore the complete soil profile.

EARLIER GENETIC SYSTEMS OF CLASSIFICATION.

The first real progress towards a valid system of classification was made by V. V. Docuchaiev² in 1879. He proposed the following classification:—

A. Normal soils (unaffected by other than pedogenic processes).

Class I. Continental humus soils.

- (a) Grey northern soils (podsoles).
- (b) Tshernosem (black earth) soils.
- (c) Chestnut earths.
- (d) Alkaline soils (solonetz).

Class II. Continental swamp soils.

B. Extra-normal soils.

Class III. Denuded or eroded soils.

Class IV. Alluvial and lacustrine soils.

This classification was somewhat modified later. It will be seen that, whilst Docuchaiev's system marks an advance, in its breadth of application, on purely physical or petrographic systems, it is only devised for the classification of soils of a limited region. No place could be found in such a system for tropical soils or for the soils of Western Europe. The inclusion of extra-normal soils in a system of classification was criticised since they are not actual soils in esse.

F. von Richtofen,³ in 1886, proposed a system based on the mode of weathering, considered from the physical and topographical standpoint. Against this classification it may be objected that it is simply a classification of regolithic materials.

E. W. Hilgard⁴ classified soils broadly into *humid* and *arid*, and showed by numerous experimental data the im-

portant differences between the groups thus distinguished. This classification, though valuable and valid, did not, however, greatly advance the solution of the general problem.

N. M. Sibirtzev⁵ was the first to throw into clear relief the general character of the factors influencing soil formation. According to him, the variety of soils is due to variations in, (a) parent material, (b) organisms participating in the soil forming processes, and (c) physico-geographical conditions, including, above all, climate. The most important factor is the humidity and, in this connexion, Sibirtzev recognised the importance of considering rainfall in relation with temperature in assessing the humidity or aridity of a region.

Three main classes were recognised, namely, A, *zonal soils*, including the succession from laterite to tundra; B, *intra-zonal soils*, including salines, peats, and rendzinas; and C, *azonal soils*, including skeletal and alluvial soils.

The ideas of Sibirtzev, though worked out in one region, Russia, mark a distinct advance towards a world system of classification: yet, with the limited material at his disposal, and with the incomplete knowledge then available as to the constitution of the soil and the nature of the pedogenic processes, it was inevitable that Sibirtzev should lack the perspective possible to later students of the subject.

E. Ramann,⁶ in his classification of the soils of Europe, recognised with Hilgard the fundamental distinction of humid and arid soils. His first classification separates the soils in which the weathering is predominantly physical as in arctic and alpine regions from those resulting from chemical weathering. The former class include tundra and mountain soils. The latter class is sub-divided as follows:—

A. Humid soils.

- (a) Podsoles, sub-divided into geographical groups distinguished by the degree of podsolisation.
- (b) Brown earths.
- (c) Yellow and red earths.

B. Arid soils.

- (a) Tshernosems.
- (b) Chestnut earths.
- (c) Grey desert, saline, and alkaline soils.

It is probably mainly to K. D. Glinka⁷ that we are indebted for our appreciation of the importance of the profile in soil studies, since his ideas have exerted the greatest influence on the present generation of students of the soil. Glinka's classification is stated in terms of maturity of profile development and the intensity of leaching by percolating waters. The following is the classification proposed:—

I. Ektodynamomorphie soils (i.e., soils in which the external factors of soil formation predominantly affect the soil character).

- (a) Laterites, red earths, and yellow earths.
- (b) Podsoles, grey forest soils and degraded tshernosems.
- (c) Tshernosems.
- (d) Chestnut earths and their related types.
- (e) Peat soils and mountain soils.
- (f) Saline and alkaline soils.

II. Endodynamomorphie soils (i.e., soils in which the parent material predominantly affects the soil character).

- (a) Rendzina, or humus carbonate soils.
- (b) Skeletal soils.

The divisions of the ektodynamomorphie soils are stated in terms of the character of the leaching which has taken place (optimal moistening, excessive moistening, etc.).

CLASSIFICATION OF SOILS ACCORDING TO CLIMATE.

Since the importance of climate as a factor in soil development has been realised, many attempts have been made to devise schemes for classification of climates, with special reference to their bearing on soil genesis. R. Lang⁸ proposes the *rain-factor*, i.e., the mean annual rainfall in millimetres divided by the mean annual temperature in

centigrade degrees, as a basis. In its later form, Lang's rain-factor is calculated for the frost-free period of the year. By his recognition that the efficiency of a given annual rainfall as a leaching agency is modified by the temperature, Lang certainly indicates an approximate method for comparing humidities. As limits for the principal soil groups Lang gives the following figures:—

Soil group	Rain-factor
Peat soils	>160
Black earths	160-100
Brown earths	100-60
Yellow earths, red earths, and laterites	60-40
Saline soils	<40

The inadequacy of this system is evident from the assignment of the black earths to a more humid climate than the brown earths, red earths, and laterites.

A. Meyer⁹ has proposed a more accurate measure of humidity in his so-called *N-S quotient*. This is obtained by dividing mean annual rainfall by a coefficient expressing the mean deficit from saturation at the mean annual temperature.

In the following table are given the N-S quotients for the principal soil groups, calculated for the whole year, and for the frost-free period, reduced to the basis of one month.

Soil groups	N-S for year	N-S for frost-free period
Desert and desert steppe ...	0-100	0-5
Mediterranean	50-200	3-18
Chestnut earths	100-275	5-10
Black earths	125-350	8-20
Brown earths	275-400	18-30
Atlantic	375-1000	25-80
Heaths	375-700	25-50
N. German & Scandinavian	300-1200	20-85
N. Russian	400-600	20-30
Alpine	1000-4000	40-350

H. Jenny¹⁰ has shown that the principal soil types in Europe and North America are characterised by definite limits for the N-S quotient.

It is obvious, from an inspection of the figures in the above table, that the principal soil groups are not defined very closely. Indeed, some of the groups distinguished are actually climatic regions. A locality with a reduced N-S factor of 25 might lie in any one of a number of groups. The classification does, however, separate the humid from the arid climates, an end which may be equally well served by Lang's rain factor.

RECENT GENETIC SYSTEMS.

During the last decade, with fuller knowledge of the soils and of the pedogenic processes occurring in different parts of the world, systems have been proposed which aim at the inclusion of all possible types. Thus, D. G. Vilensky¹¹ distinguishes four broad divisions of soils, based on the dominant factors in their formation. These are (1) *thermogenic*; (2) *phytogenic*; (3) *hydrogenic*; and (4) *halogenic*.

(1) Soils of the thermogenic divisions are developed in sub-tropical and equatorial regions, in which the dominant factor is the high temperature, which causes (a) rapid chemical decomposition of mineral silicates, and (b) rapid mineralisation of plant residues with production of carbon dioxide. Under these conditions, red and yellow loams and laterites result.

(2) Soils of the phytogenic division are mainly developed in the temperate zones under a wide range of humidity. They are characterised by conditions which favour the accumulation of organic matter in the soil, and involve a less intense weathering of mineral silicates than in the thermogenic divisions. These soils include the tshernosem and chestnut earth group, the degraded tshernosems, and the podsols.

(3) Soils of the hydrogenic division are formed chiefly in cold climates, i.e., in the tundra and the adjacent forest regions. Soil formation proceeds mainly under water-logged conditions with the development of peaty humus. Ferrous compounds such as pyrites, marcasite, and ferrous carbonate, are found in the sub-aqueous horizons. The soils include the tundra soils, the peat podsol, and the meadow soils.

(4) The halogenic division includes soils developed in the presence of sodium salts, and includes the saline, alkaline, and soloti soils.

Intermediate divisions such as thermophytogenic and thermohydrogenic are also distinguished.

Vilensky¹² has more recently developed and re-arranged his system in the following form, which throws into clear relief the relative parts played by temperature and humidity in soil formation.

D. G. VILENSKY'S SYSTEM OF CLASSIFICATION.

	Arid	Semi-Arid	Medium	Semi-Humid	Humid
Polar ...	Tundra	Semi-peat soils	Peat and meadow soils	—	Podsolised peat and meadow
Cold	Dry peat soils	—	Black meadow soils	Degraded meadow soils	Podsolised soils
Temperate	Grey earths	Chestnut earths	Tshernosem	Degraded (grey forest) soils	Podsolised soils
Sub-tropical	—	Yellow soils of dry steppe	Yellow earths	Degraded yellow earths	Podsolised yellow earths
Tropical ...	Red soils of semi-desert	Red earths	Laterite	Degraded red earths	Podsolised red earths

S. S. Neustreuev,¹³ in an attempt to classify soils according to the processes by which they have been formed, distinguishes *hydromorphous* and *automorphous* processes.

Hydromorphous processes are those which take place

under the influence of ground-water, which directly, or by capillary action, moistens the soil profile. In automorphous processes, ground-water influence is absent.

The hydromorphous decompositions are divided into two types. In the one type, capillary transport is dominant, giving rise to saline soils, and meadow soils with turf or bog iron-ore formation. In the second type, anaerobic conditions dominate profile development and give rise to meadow and peat soils.

The automorphous processes are divided into three types depending on intensity of decomposition of the mineral silicates. In the most intense type, complete decomposition into silicic acid, sesquioxides, and aluminosilicic acids takes place. Differentiation within the profile gives the subdivisions, laterite, bauxite-laterite, and kaolinite-laterite. A less intense variety of this type is found where intermediate products of decomposition also occur. According to the type of eluviation, red, yellow, and brown earths are produced on the one hand, and podzols and solochs on the other.

In the second type of decomposition, free sesquioxides are not formed and the principal products are acid aluminosilicic acids (clays with high $\text{SiO}_2/\text{R}_2\text{O}_3$ ratio). This type, formed under semi-humid, semi-arid and arid conditions, gives rise to tchernosems, chestnut earths, and grey-brown or grey steppe soils.

In the third type of automorphous decomposition, there is only a slight chemical weathering. Products of physical weathering predominate, whilst the fine material, if present, is mainly of external origin (wind or water transport). The soils produced by this type of decomposition include arctic and alpine soils on the one hand, and desert soils on the other.

A. Stebut¹⁴ has developed a system of classification also based on the character or tendency of the pedogenic processes. According to this author, the three principal groups of processes in soil formations are:—(1) Decomposition, affecting silicate minerals (hydrolysis). (2) Synthesis

of new products (formation of the weathering complex).
 (3) Differentiation into horizons (eluviation and illuviation).

After reviewing these processes in detail, Stebutt distinguishes:—

A. UNDEVELOPED SOIL FORMATION.

Class I. Lack of development due to external factors, i.e., low temperature or aridity.

Class II. Lack of development owing to nature of parent material, e.g., quartzite or quartz sand.

Class III. Development potential but unaccomplished, e.g., recent alluvia.

B. DEVELOPED SOIL FORMATION.

Class I. Zeolith (Clay) formation.

Sub-class (a) In presence of alkali salts (saline and alkaline soils).

Sub-class (b) After dealkalisation (tshernosems, etc.).

Class II. Degradation.

Sub-class (a) Degradation of alkali soils (soloti).

Sub-class (b) Degradation of calcareous soils (degraded tshernosems).

Class III. Destruction.

Sub-class (a) Acid humus destruction.

i. Oxidative conditions.

α . No superficial humus accumulation (brown earths).

β . Formation of peat layer (podsoles).

γ . Raw humus accumulations (peat podsoles).

δ . Epigenetic peat formation (peat).

ii. Reducing conditions (peats and meadow soils).

Sub-class (b) Destruction by carbonic acid.

α . Without eluviation (red earths and brown earths).

β . With eluviation (laterites).

The above scheme gives a comprehensive view of the pedogenic processes and seems capable, by elaboration, of including all possible types and varieties of soil. In view of the importance of the character of the parent material in modifying the trend of the pedogenic processes, systems of classification such as those of Neustreuev and Stebutt mark a decided advance on systems which are stated in terms of climate alone.

The systems hitherto discussed have been based on a consideration of the pedogenic processes. An attempt to classify soils according to the character of the absorbing complex has been made by K. K. Gedroiz,¹⁵ who first distinguishes, (1) soils saturated with bases; and, (2) soils unsaturated with bases. The base-saturated soils fall into (a) the tshernosem group, predominantly saturated with calcium and magnesium, and (b) the alkaline group in which sodium is dominant, including the saline, alkaline, and soloti soils. The base-unsaturated soils fall into two groups, namely, (a) the podsoles and, (b) the laterites, including red and yellow earths, varying in the extent to which the weathering complex is decomposed.

Although Gedroiz's system is valid in itself, it seems scarcely detailed enough in its present form to serve as a world system.

During the past decade, considerable discussion has been directed towards the institution of an agreed world system of soil classification. C. F. Marbut¹⁶, at the 1st International Congress of Soil Science, held at Washington in 1927, proposed a system of soil classification to include all soils already known or likely to be encountered.

Marbut begins by making a division of soils into *pedocals* and *pedalfers*. Pedocals, which may be regarded as synonymous with Hilgard's arid soils, are characterised by the presence in the soil profile of a zone of calcium carbonate accumulation. Pedalfers, synonymous with humid soils, show no accumulation of calcium carbonate in the soil horizons, but show a differentiation or tendency to

differentiation of the clay complex in the different soil horizons, resulting in accumulation of sesquioxides.

In the next stage* of classification, the pedocals are divided into pedocals of the temperate zone and pedocals of the tropical zone, whilst the pedalfers are divided into podsollic soils and lateritic soils. The basis of classification in the second stage is, thus, temperature.

In the third stage of classification, the pedalfers are divided into the following groups: (1) tundra; (2) podsoles; (3) brown forest soils; (4) red soils; (5) yellow soils; (6) prairie soils; (7) laterites; and (8) ferruginous laterites. The pedocals are similarly distinguished into, (1) north-temperate pedocals; (2) mid-latitude pedocals; (3) south-temperate pedocals; and (4) tropical pedocals. It will be seen that, in this stage of classification, the distinctions made in the second stage are included.

In the fourth stage of classification, distinctions based on rainfall are used. In the scheme, as outlined by Marbut, this is only applied to the mid-latitude pedocals, giving the succession: tshernosem; chestnut earths; brown (desert) earths; and grey (desert) earths. The information at present available does not permit a similar differentiation in the case of pedocals of warmer and colder climates. It is considered that this basis of differentiation may be applicable also to the pedalfers.

In the fifth stage of classification, soils are differentiated on the basis of the degree of maturity of the soil profile.

In the sixth stage of classification, distinctions based on the parent material are used, giving the soil series, whilst in the seventh stage the texture is used, giving the type.

Marbut's classification is shown schematically in Table XXXVIII. It will be seen that spaces are left for the insertion of groups still to be identified and described, so that, with some necessary modifications, the scheme does give a basis for a world wide system of classification. It will be

* The writer prefers the term *stage* to *category*, used by Marbut, and has ventured to introduce this emendation in the present account.

noticed, however, that no place is found for soils developed under conditions of impeded drainage, such as meadow and vlei soils. Such soils being intra-zonal or azonal, may not be considered as demanding a place in the world-system, on account of their dependence on special local conditions of topography and drainage. They are, nevertheless, of widespread occurrence and, in the writer's opinion, should form a major group. Neustreuev, indeed, thus recognises them in his class of hydromorphous soils.

TABLE XXXVIII. — SCHEME OF SOIL CLASSIFICATION BY C. F. MARBUT.

Stage 1	Stage 2	Stage 3	Stage 4	Stage 5	Stage 6	Stage 7
PEDOCALS	Pedocals of Temperate Climates	North- Temperate		Subdivision of categories of Stage 4 according to maturity of profile	Subdivision of categories of Stage 5 according to parent material	Subdivision of categories of Stage 6 according to surface texture.
		Mid-Latitude	Tshernosems Chestnut- Earths Brown Soils Grey Soils			
		South- Temperate				
PEDALFERS	Pedocals of Tropical Climates			Subdivision of categories of Stage 4 according to maturity of profile	Subdivision of categories of Stage 5 according to parent material	Subdivision of categories of Stage 6 according to surface texture.
	Podsolie Soils	Tundras				
		Podsols				
		Brown Forest Soils				
		Prairie Soils				
	Lateritic Soils	Red Earths				
		Yellow Earths				
		Laterites				
		Ferruginous Laterites				

SIGNIFICANCE OF CLIMATE, GEOLOGY, AND
TOPOGRAPHY.

Mature soil profiles are the result of the pedogenic processes acting on different materials, which may be either consolidated rock, or unconsolidated *débris*, such as alluvium or loess. In arriving at a classification of profiles, it is of importance to obtain a clear view of the relative weight to be assigned to the pedogenic processes on the one hand and to the parent material on the other as factors in soil development.

In the earlier systems of soil classification, the dominant factor was considered to be geology, as affecting the parent material. The studies of the Russian school of pedologists, with their emphasis on climatic factors, have led to greater weight being assigned to the pedogenic processes, which were considered to be capable of obliterating the differences due to parent material. It was found, for example, that black earth profiles could be developed from parent materials which differed considerably in lithological character. On the other hand, instances were brought forward in which the same parent material gave rise to totally different soils under the action of different pedogenic processes. A classical example is that of two dolerites of approximately the same composition, one of which yielded under English conditions a grey clay, whilst the other, in Southern India, gave rise to a laterite, consisting almost entirely of hydrated aluminium and ferric oxides.

Since the pedogenic processes are mainly governed by temperature and by the balance of rainfall and evaporation, it is natural that systems of classification should be elaborated which are essentially climatic. Indeed, the criticism might be made that such systems are classifications of climate rather than of soil. Even in classifications, such as that of Marbut, which meet this objection and purport to be based on the actual soil characters, it is impossible to avoid defining the classes in terms of climate.

Now, whilst it is certain that the major differences

between soils are due to the effect of climate operating through the pedogenic processes, the geological factor cannot be eliminated even in the definition of the great soil groups. Many instances may be given in which, under a similar set of pedogenic processes, different parent materials have given rise to soils which must be assigned to different world groups.

B. Polynov,¹⁷ in a discussion of the rôle of geology in soil formation and its significance in classification, directs attention to the effect of parent material in modifying the occurrence and distribution of the great soil groups. Quartzose sand enlarges the area of podsoles both at the northern and southern limits. Sandy parent materials also increase the desert area. For example, in the Caspian region, whilst loam soils belong to the chestnut earths, sandy soils have a desert character.

In South-Eastern England, typical podsoles are developed on the light sands and gravels of the Bagshot Beds. Under the same climate, other parent materials yield brown earths and even soils which may have affinities with terra rossa. It may be said of Britain as a whole that the lightest sands form podsoles, whilst parent materials of heavier texture and higher base status yield brown earths.

In the Transvaal, the weathering of diabase yields red loams which are probably related genetically with the tropical red loams, whilst norite gives rise to black soils, the so-called black turf soils, which resemble, in some respects, vlei soils and must certainly be assigned to a major group distinct from the red loams.

W. H. Bryan and H. J. G. Hines¹⁸ have reported even more striking instances of variety of soil type under the same climate in Queensland. There, within a comparatively small area of approximately uniform climate, soils comparable with the tshernosems, podsoles, red earths, yellow earths, and laterites of other countries are to be found. This diversity can be satisfactorily explained by considerations of parent material and topographical relief.

Soils derived from calcareous parent materials exhibit certain distinctive features. On the one hand are the red soils of the terra rossa group, whilst on the other hand are the grey or black soils of the rendzina group. It appears that, so long as the calcium carbonate of the parent material can maintain a state of base-saturation, the soil is of a greyish colour with a weathering complex relatively rich in silica and a dark-coloured type of humus. But when desaturation occurs through leaching—a state which may be reached before the complete disappearance of coarsely fragmental limestone—desilicification ensues and reddish or reddish-brown soils are formed, containing free sesquioxides in the weathering complex. This stage is reached most readily in soils derived from hard limestones, which thus tend to be associated with reddish soils, whilst soft limestones generally give rise to greyish or dark-coloured soils of the rendzina group. The ease with which leaching occurs in hard limestone soils, the rapid mineralisation of plant residues consequent on their warmth and aeration, all contribute to confer on such soils the characters of soils of warmer climates. Mature leached limestone soils of sub-tropical and even of temperate regions show considerable resemblance to tropical red loams.

It is obvious, therefore, that a classification of climates without reference to geology can only give a partial classification, even of the major soil groups. A map showing the major soil groups must in many cases show different groups occurring under the same climatic conditions.

In the more detailed classification of the soils of a limited region, geology is frequently the main guide. After all due weight has been given to climatic factors, the most fruitful principle of classification of British soils is the geology, if consideration be given to lithological variations.

Even in a region of uniform climate, geology does not always provide a sufficient basis for differentiation, for important modifications are introduced by topography. Some of these modifications are in a sense geological. The differ-

ence between the thin soils of the uplands and the deep colluvial soils of valleys may be expressed in terms of superficial geology. In a detailed soil survey, such distinctions as *shallow phase* and *deep phase* are of great importance.

Topography exerts its principal effect on soil development by its bearing on water movements. The same parent material gives rise to markedly different soil profiles according to whether the drainage is free or impeded. With free drainage under humid conditions, the soil is more strongly leached of bases than where the drainage is impeded. This leads to degradation of the clay complex with liberation of silicic acid and sesquioxides. In cool climates, an acid type of organic matter is formed from the residues of vegetation, and a podsol profile results. The brown earths represent an intermediate stage, but seem to be essentially similar to the tropical red loams.

In low-lying situations with a high water table and impeded drainage, the profile development will depend upon the extent to which the accumulation of acid organic matter occurs. In regions of high rainfall and moderate temperature, a peaty layer is formed and a bleaching of the mineral soil through removal of sesquioxides occurs, resulting in a peat podsol. This process may, indeed, occur in water-logged situations in the tropics.

Where peat is not formed, meadow soils occur in cool and temperate climates. These are characterised by grey horizons with rusty mottling in the vicinity of the fluctuating water-table. In similar situations in the tropics, black and grey soils are found, also with rusty mottling, and even concretionary deposits, in the vicinity of the water-table.

The mineral colour of soils with impeded drainage is generally grey owing to the presence of ferrous compounds consequent on deficient aeration. Further, the low degree of leaching and, in many cases, the relatively high base status result in the development of a clay complex relatively rich in silica.

The occurrence of saline and alkaline soils can generally

be attributed to the effect of topography operating through drainage conditions.

The relative weight to be assigned to the factors geology, climate, and topography, will depend on the circumstances of a particular region or country. In Russia, for example, climate has proved the most fruitful principle of classification. In Great Britain, geology gives the clearest clue to soil differences. In every region, however, topography, as affecting the drainage and as modifying the depth of accumulation of the products of weathering, enters as an important factor in classification. The principle to be observed in every case is to classify on the basis of the soil itself, using the profile as the unity of study.

PROVISIONAL CHARACTER OF EXISTING SYSTEMS OF CLASSIFICATION.

The attempts which have been made to devise a world classification of soils have been encouraged by the impression that identical, or nearly identical, soils can occur in different parts of the earth. Thus, it is generally considered that, in the high plains of the United States, there are soils which can be classed with the tshernosem and related groups of Russia. In view of the qualitative character of much of the information relating even to such an important group as the tshernosems, it would be safer to regard such correlations as provisional, and to await the accumulation of such a body of quantitative data as shall place their identity or dissimilarity beyond doubt.

This view does not imply that the proposed systems of classification are not worthy of serious consideration. On the contrary, so far as they are based on a considered review of intensely studied soil types, they are of the highest philosophical value. Their provisional character consists in the uncertainty of their application to soils which have not yet been subjected to intensive study, or which have only been described in qualitative and, therefore, subjective terms.

And thus, whilst the various systems proposed may be trustworthy for the soils on which they are based, and suggestive for soils yet to be studied, the immediate task is to enlarge and intensify our information as to the constitution and genesis of the soils of different parts of the world. And if, as is probable though not proven, the soils of the world can be grouped into a finite number of classes, then the modification of existing systems and the synthesis of local classifications will ultimately lead to a world system.

For these reasons, the provisional character of all methods of soil classification hitherto proposed cannot be too strongly emphasised. There are large tracts of the earth's surface where soils have been so imperfectly investigated that no trustworthy conclusions can be drawn as to their relationships. Even for the better known regions, the data are frequently of a qualitative character, and the deductions therefrom may be fraught with serious errors.

Whilst the pedologist should always strive to relate his soils to those already described in other countries or regions, he should not adopt uncritically types recognised under different conditions of topography, geology, and climate. He should, in the first place, distinguish the different profiles in the region for which he is responsible. Each profile should be described with the fullest qualitative and quantitative data, including information as to mechanical composition, exchangeable base content, salt content, and the composition of the clay fraction in each distinguishable horizon. In addition, data should also be obtained for the organic matter profile, making use of any methods of fractionation which may become available. With the aid of such data, the profile types should be arranged into a system, which can then be related to systems recognised in other countries.

It is along these lines, in the writer's opinion, that a satisfactory world system of classification will finally be reached. It is true that in some cases it will be possible to identify a soil type in one country with a well known soil

type already recognised in another country, but the factors of soil formation and the parent materials on which they operate are so diverse that it must be an exceptional case for a soil type isolated, say, in Europe, to be completely identifiable with a type in North America or Australia. Premature identification may be a source of irreparable error.

The most fundamental distinctions, which seem likely to survive any changes due to the widening of our knowledge, appear to be those based on the character of the leaching processes. We may thus distinguish:—

I. Soils with complete leaching, including (a) podsollic soils, developed under the influence of acid humus; and (b) brown earths, yellow earths, red loams, red earths, and some types of laterite, developed in the absence of acid humus.

II. Soils with impeded leaching, including meadow soils, vlei soils, and the group which includes saline and alkaline soils.

III. Soils with incomplete leaching, including the succession from black earths to desert soils.

Group I corresponds with the pedalfers and Group III with the pedocals distinguished by Marbut. Groups I and III are the automorphous, and Group II the hydromorphous soils of Neustreuev.

BIBLIOGRAPHY.

- ¹ FALLOU, F. A., *Pedologie oder allgemeine und besondere Bodenkunde*, Dresden, 1862, pp. 180-182; from K. D. GLINKA, *Die Typen der Bodenbildung*, Berlin, 1914, pp. 18-19.
- ² DOUCHAIEV, V. V., *Arbeit St. Petersb. Naturforschergesellsch.*, 1879, **10**; from GLINKA, pp. 21-22.
- ³ RICHTOFEN, F. VON, *Führer für Forschungsreisende*, 1886; in GLINKA, pp. 23-24.
- ⁴ HILGARD, E. W., *Soils*, Macmillan Company, New York.
- ⁵ SIBIRTZEV, N. M., in GLINKA, pp. 27-28.
- ⁶ RAMANN, E., *Bodenkunde*, Berlin, 1911, pp. 578, *et seq.*
- ⁷ GLINKA, K. D., *Die Typen der Bodenbildung*, Berlin, 1914.
- ⁸ LANG, R., "Versuch einer exakten Klassifikation der Böden in klimatischen und geologischen Hinsicht." *Int. Mitt. Bodenk.*, 1915, **5**, pp. 312-346.

- ⁹ MEYER, A., "Über einige Zusammenhänge zwischen Klima und Böden in Europa." *Chem. der Erde*, 1926, pp. 290-347.
- ¹⁰ JENNY, H., "Klima und Klimabodentypen in Europa und den Vereinigten Staaten von Nordamerika." *Soil. Res.*, 1929, **1**, pp. 139-189.
- ¹¹ VILENSKY, D. G., "The classification of soils on the basis of analogous series in soil formation." *Proc. 1st Int. Soc. Soil Sci.*, New Series, **1**, pp. 224-241.
- ¹² VILENSKY, D. G., "Concerning the principles of a genetic soil classification." *Contributions to the study of the soils of Ukraina*, 1927, **6**, pp. 129-151.
- ¹³ NEUSTREUEV, S. S., *Bull. Geogr. Inst.* (Russian), 1926; in J. N. AFANASIEV, "The classification problem in Soil Science." *Russian Pedological Investigations V.*, Academy of Sciences, Leningrad, 1927.
- ¹⁴ STEBUTT, A., *Lehrbuch der allgemeinen Bodenkunde*, Berlin, 1930, pp. 293, *et seq.*
- ¹⁵ GEDROIZ, K. K., "Der adsorbierende Bodenkomplex und die adsorbierten Bodenkationen als Grundlage der genetischen Boden-Klassifikation." *Kolloidchem. Beihefte*, 1929, pp. 1-112.
- ¹⁶ MARBUT, C. F., "A scheme for soil classification." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 1-31.
- ¹⁷ POLYNOV, B., "Das Muttergestein als Faktor der Bodenbildung und als Kriterium für die Bodenklassifikation." *Soil Res.*, 1930, **2**, pp. 165-180.
- ¹⁸ BRYAN, W. H., and HINES, H. J. G., "Factors in the development of soil profiles in Southern Queensland." *Soil Res.*, 1931, **2**, pp. 264-276.

CHAPTER XVI.

THE GEOGRAPHY OF SOILS.

IN the present state of our knowledge, it is, perhaps, hazardous to attempt to give any general view of the soils of the world, for large areas remain pedologically unexplored. Such information as can be given about the soils of these regions is, to a large extent, conjectural, and based on the collateral evidence of climate, topography, and vegetation. Nevertheless, a review of the soils of the world, in the light of such knowledge as we possess, is desirable, and will be attempted, with all reserve, in the present chapter.*

It is only to be expected that fuller information is available in some countries than in others. The lack of accurate data is most marked in tropical regions, where the material hitherto available has consisted mainly of analyses of soil samples made for practical requirements. Only during recent years have descriptions of actual tropical soil profiles appeared in the literature of the subject.

EUROPE.

As may be expected, the information concerning soil is more abundant in Europe than in any other portion of the earth, with the possible exception of the United States; but there are still considerable gaps in our knowledge. A large amount of material is available relating to the soils of Russia, Poland, Scandinavia, Holland, Germany, the former Austro-Hungarian Empire, and Great Britain. Considerably less material exists for judging the soils of France, Spain, Portugal, Italy, and the Balkan Peninsula.

At the instance of Commission V of the International Society of Soil Science, a soil map of Europe has been

*For the sake of convenience in the narrative, the sources are not given in the text, but are set out at the end of the chapter.

prepared under the direction of Professor H. Stremme, of Dantzig. A provisional map on the scale of 1:10,000,000, with an accompanying memoir, was published in 1927, and a revised map on the scale of 1:2,500,000 is now in active preparation. Making allowance for the somewhat Procrustean system of classification and for the uncertainty as to the position of the soils of certain regions in this system, it is yet possible to make certain generalisations as to the distribution of soils in the different countries.

Great Britain. The soils of Great Britain belong mainly to the podsolic group, if in this group we may include the brown earths. In the provisional soil map of Europe, most of the eastern and southern part of Britain is shown as brown earths, with podsols lying to the north and west. Recent experience has shown the important part played by topography and, above all, by the parent material in determining the character of British soils. The tendency to podsolisation becomes more pronounced in moving from south-east to north-west, but quartzose sands in S.E. England carry well developed podsols, whilst, even in the most humid regions of the west, brown earths occur where the parent materials are such as to yield soils of intermediate or heavy texture. Further, the relative immaturity of much of our British soil results in the parent material playing a great part in determining actual soil characters.*

Such being the importance of geology, there appears to be ample justification for reviewing the soils of Britain in terms of the parent materials which have produced them.

Pre-Cambrian rocks attain their greatest extent in the Scottish Highlands, where they are represented by schists and sandstones. In the cultivated valleys and foothills, the soils, though generally immature, appear to be usually of a brown earth type where well drained, whilst, in upland areas, podsols are more prevalent. With them are associated meadow soils, peat podsols, and peats. The most

*Age-long cultivation has probably produced considerable modifications, so that soils which were originally podsols or meadow soils have now become brown earths.

considerable area of Pre-Cambrian rocks in Southern Britain is in Anglesey, where brown earths are found derived from schists and schistose drifts, with occasional patches of iron podsols in uplands, particularly in association with quartzites.

There is a general similarity between soils derived from igneous and pyroclastic rocks of all periods. Further, with the relative immaturity of British soils, the differences between the weathering of basic and acid rocks are not so marked as in warmer climates. Generally speaking, igneous rocks give rise to brown earths in lowland areas and podsollic soils, sometimes truncated by erosion, in uplands. The soils of the more acid rocks, such as granite and rhyolite, more readily undergo podsolisation than those derived from igneous and basic rocks.

The sedimentary rocks of Cambrian, Ordovician, and Silurian age (excepting hard grits, which appear to behave pedogenically as igneous rocks) give rise, in lowland areas, to soils which are generally similar in character and may be classed together as brown earths. But, since they occur mainly in the west under high rainfalls, drainage impedance is frequent and considerable areas of meadow soil result. The uplands are occupied by shallow podsols, often truncated by removal of the A horizon so that the surface soil is of a markedly sesquioxidic character. Meadow soils, peat podsols, and mountain peats are of frequent occurrence. The soils of the Devonian shales in Devon may also be grouped with those of the older Palæozoic sediments.

The Old Red Sandstone gives rise mainly to brown earths, but some of the quartzose conglomerates carry podsols, as in W. Glamorganshire. The soils of the uplands exhibit various stages of podsolisation, modified by erosion. The bright red colours of certain soils of this formation are to be attributed to the effect of the parent material.

The character of the soils of the Carboniferous Limestone is greatly influenced by topography. The colluvial and glacial drift soils may have affinities with the rendzinas.

On the other hand, there are thin leached upland soils which may be assigned to the group of red and brown soils to which terra rossa belongs. In some cases, soil formation is almost entirely absent, and limestone pavements result.

The non-calcareous sedimentary rocks of the Carboniferous system give rise mainly to brown earths, but heavy texture sometimes results in drainage impedance and the development of meadow soils. In Wales, the Millstone Grit gives greyish soils which have a highly siliceous type of clay. They do not appear to be podsolised, for the grey colour often persists to the rock without any sign of a B horizon. Similar soils are developed from certain siliceous sandstones of the Cretaceous system. The siliceous character of the clay is doubtless attributable to the nature of the parent material, rather than to the operation of contemporary processes.

The Permian formation in England is represented by the Magnesian Limestone, whose associated soils resemble, in many respects, those of the Carboniferous Limestone, and by breccias, marls and sandstones, whose soils may be grouped with those of the Trias.

The Trias may be considered briefly under the headings, sandstones and marls. The sandstones, which are generally red in the Bunter, but sometimes brown or grey in the Keuper, give rise to brown earths and podsoles. The distinction appears to depend on the actual character of the material, the more quartzose sandstones and the pebbly conglomerates, with their associated drifts, being most apt to become podsolised. Drainage conditions, by their effect on leaching, also play a part. Meadow and peat soils are of frequent occurrence in certain drift areas of Cheshire and Shropshire.

The Triassic (Keuper) marls are mainly reddish clays, sometimes containing calcium carbonate, which, however, is generally leached from the surface horizons. The derived soils are brown earths and meadow soils.

Bright red soils occur frequently in association with cer-

tain upper horizons of the Carboniferous system, the Permian sandstones, and the Triassic sandstones; but it should be clearly understood that the colour is due to the character of the parent material and is not attributable to contemporary pedogenic processes.

In considering the Jurassic and later systems, with their rapidly alternating succession of sediments, we shall find it convenient to consider them from the point of view of lithology, which, subject to topographical factors, appears to supply an adequate classification.

Soft limestones, such as the Lower Lias and parts of the Chalk, give rise to greyish-brown, greyish, and in some cases, white soils, which may be grouped with the rendzinas. In situations where calcium carbonate is leached away, they may give brown or reddish-brown soils which are essentially brown earths.

Hard limestones, such as the Forest Marble and the Kentish Rag, may give reddish-brown soils akin to the brown earths and, possibly, to terra rossa. Where base-saturation is maintained, the grey or grey-brown rendzina type persists.

Clays, such as the Oxford, Kimmeridge, and Gault clays, generally suffer drainage impedance, and their associated soils may be grouped with meadow soils. Horizons of calcium carbonate or gypsum accumulation sometimes occur with drainage impedance. With free drainage, brown earths are formed.

The sandstones of the Jurassic and later formations are, for the most part, soft or unconsolidated. The derived soils form a range varying from brown earths to podsoles according to the lithological character of the parent material. Quartzose sands and gravels, such as the Bagshot Beds, form well-developed podsoles with massive humus and iron pans.

The nature of the soils derived from glacial drifts is generally governed by the character of the contributing materials and by the topography and drainage.

Alluvial soils occur in all parts of Britain. They are generally immature, except in the case of old river terraces, where in some cases a certain amount of profile development may have occurred. Dune soils in varying stages of fixation occur in coastal districts.

The most notable development of lowland peat is in the Fen district. Upland and mountain peats occur commonly in elevated areas, particularly in the west.

Much of the soil of upland areas in Britain is very immature and may be described as skeletal. It is probable that erosion has produced considerable modification, so that the shallow profiles encountered are considerably truncated, and only represent the lower horizons of the original profiles existing under the primitive forest cover, which extended up to about 1,500-1,700 feet above the sea level.

Scandinavia and Finland. A considerable proportion of Scandinavia is occupied by immature mountain soils with podsollic tendencies. The lowlands of Scandinavia and Finland are occupied by glacial deposits, including boulder clays, lacustrine clays, and fluvioglacial sands and gravels. Podsoles, in varying stages of development, form the typical soil group of these regions and are developed naturally under birch forest, coniferous forest, or in Denmark, heath. Brown earths occur in parts of Southern Scandinavia under deciduous forest. The topography and hydrological conditions associated with a region of glacial drifts result in the widespread occurrence of peats and meadow soils in low-lying situations. In an area of low rainfall in Central Norway, saline soils are encountered.

Poland. Podsollic soils in varying degrees of maturity occur in Poland. An extensive area of meadow soils is found in the Pripet marsh region. Rendzinas are also present, whilst tshernosems and degraded tshernosems are found in the south.

Germany. Podsoles and their congeners occupy the greater part of the area. Varying degrees of podsolisation occur from brown earths to strongly podsolised soils. Areas

of tshernosems and degraded tshernosems occur in Central Germany in isolated patches. Immature skeletal soils are found in mountainous regions in the south, where podsolised soils and meadow soils also occur locally.

Calcareous parent materials give rise to rendzinas, and, in some cases, reddish-brown soils.

Meadow soils, peat soils, river alluvia, and estuarine alluvia occur as topographical types.

Czechoslovakia, Austria, and Hungary. A wide range of soils is shown in these countries. In the mountainous regions, there are immature skeletal soils generally of a podsollic type, but including some rendzinas. Among mature soils of the lowlands the brown earths predominate, but the Central Hungarian plain carries dry steppe soils with associated saline and alkaline (szík) soils. Red soils occur in Moravia and Hungary (nyírok soils) which appear to have affinities with terra rossa but may have resulted from Tertiary weathering.

Italy and the Adriatic coastal lands. The soils of northern Italy and the western coastlands of the Adriatic are mainly brown earths. Elsewhere in Italy, including Sicily, skeletal soils, markedly affected by erosion, prevail. The tendency, however, is towards the development of red soils, which may probably form a transition between the brown earths and the tropical red loams and red earths. Terra rossa is developed on limestone in the northern and eastern coastal regions of the Adriatic.

France and Belgium. This is largely a region of brown earths, but the chalk soils of northern France may be assigned to the rendzinas. Podcols occur in Brittany, whilst, in the south, terra rossa associated with limestone, and yellow earths occur.

Holland. A large proportion of this country is occupied by estuarine alluvium reclaimed from the sea. The soils of the older reclamations may possibly be assigned to the meadow soils and brown earths. Podcols and upland peats occur in northern Holland.

Spain. Comparatively little work has hitherto been carried out on the soils of Spain. The available data are based principally on ecological studies. Whilst the West European and Mediterranean soil groups are represented, there is one group of reddish-brown soils which appears to be specially characteristic of Spain. These soils occur in Central Spain under semi-humid conditions, associated with xerophytic forest, and are termed by E. del Villar *calvero* soils. The natural profile has been somewhat modified by erosion consequent on human interference. Their affinities are not at present clear, but the character of the natural vegetation, the low humus content, and the general leaching of carbonates, preclude their identification with the pedocals. They may, indeed, prove to have affinities with the red loams found under savanna and monsoon forest in the tropics.

Del Villar describes certain dark soils, which may possibly be meadow or vlei soils, whilst saline and alkaline soils occur in Murcia.

Balkan Peninsula. On account of the strong relief of this region there is generally a marked vertical zonation of soils. Brown earths, red soils resembling the nyirok soils of Hungary, and terra rossa are found in lowland regions, whilst podsolised and skeletal soils occur at higher elevations. It would appear that considerable modifications have been produced by erosion, particularly in Greece, where the deep stoneless clays of the plains are sharply contrasted with the thin stony soils of the uplands. The tradition of a former cover of soil carrying forest in the uplands of Greece is mentioned in Plato's *Critias*.

Russia. Modern conceptions of soil genesis have largely developed from the studies made on the soils of Russia. The dominance of Russian ideas in systematic pedology is the natural outcome of the co-ordinated efforts of a national school of investigators having as its province the wide range of climatic conditions implied by the vast extent of the Russian Empire.

The succession of soil groups in European Russia is relatively simple. In the extreme north is the tundra region, succeeded to the south by a broad belt of podsoils which pass into the degraded tshernosems or grey forest soils.

The boundary separating the grey forest soils from the tshernosems runs roughly south-west to north-east, passing through the Ukraine. The tshernosem zone is succeeded to the south-east by the chestnut-coloured earths, and these, in turn, pass with increasing aridity into the grey and brown semi-desert soils of the Caspian region. Saline and alkaline soils occur as local developments in the Caspian region.

ASIA.

*India.** The most salient feature of the soils of India is the large proportion of alluvium. The greater part of Northern India is occupied by the deposits of the Indus, Ganges, Brahmaputra, and their tributaries. Whilst recent alluvium cannot be expected to reflect the operation of the pedogenic processes consequent on the local climate, the older alluvia show varying degrees of maturity. There is thus a marked distinction between the soils developed under the semi-arid climate of Rajputana and the Punjab, and those developed in the more humid climate of Bengal. Soils with carbonate accumulation and soils of the saline and alkaline group occur in the former areas, but are absent from Bengal.

The southern part of the peninsula is occupied partly by red earths and laterites, the latter occurring along the western coast and in Bihar, Orissa, and Chota-Nagpur, and partly by black cotton soils.

The black cotton soils, termed *regur* in Central Provinces and Hyderabad, occupy an area lying to the east of the Western Ghats and including a large portion of these Provinces. The affinities of the black cotton soils with the black earths have been discussed in Chapter X. Typically they are derived from trap rock—the so-called Deccan trap.

*The writer is indebted to Dr. H. H. Mann, of the Woburn Experimental Station, for much of the information on which this section is based.

The depth varies considerably: it may reach as much as twenty feet in areas of accumulation, but a depth of one to five feet is more usual.

When wet, the black cotton soils form a slimy colloidal mass, but on drying out they readily pulverise so that the soil is said to "plough itself." The mechanical analyses of certain typical soils do not, however, show high proportions of clay, and the reason for their highly colloidal character calls for investigation.

The development of the zone of carbonate accumulation varies somewhat and may in some cases be almost vestigial. It is possible that some of the black cotton soils may prove to have affinities with the rendzinas on the one hand and the vlei soils on the other, whilst an analogy with the Transvaal black turf soils may also be suggested. It is significant that in certain areas where black soils occur, the higher ground is occupied by red soils, which may stand in the same relationship to the black soils as the red limestone soils to the black limestone soils in Barbados.

Comparatively little information is available as to the red soils which occupy most of southern India outside the black cotton area. Developed under a monsoon climate and mainly derived by primary weathering from crystalline rocks, they are probably similar to the red loams and red earths of East Africa. Soils which may be analogous to the chestnut-coloured earths occur in Hyderabad to the east of the black cotton area.

A considerable area of desert and semi-desert occurs in Sind and Rajputana.

Siberia. In Glinka's "Typen der Bodenbildung" a soil map of the Russian Empire on the scale of 1:20,000,000 is given. The zonation of soil groups in Siberia follows generally the parallels of latitude, and the succession from north to south is from tundra through podsoles, black earths, chestnut earths, to grey desert soils. The podsol region extends further south in the mountainous regions adjoining Thibet and China. Most of the area above 60°N latitude is

tundra. Degraded tshernosems, saline and related soils, and desert soils occur in Turkestan.

Asia Minor. Recent studies have shown the occurrence of terra rossa in Palestine. Comparatively little is known of the soils of Syria and Asiatic Turkey, but they probably include intermediate stages between the brown earths of Europe and the red soils of the tropics. In the drier regions, the black earth to chestnut earth succession probably occurs. Podsoles and skeletal soils may be expected in mountainous regions.

China. The available information on the soils of this country has been mainly in connexion with descriptions of Chinese agriculture. Recently, however, a preliminary account by C. F. Shaw of the soils of China from the standpoint of modern pedology has been published by the Geological Survey of China.

A large proportion of the area is occupied by immature alluvial soils, which are generally deep silt loams and clays. Brown earths, and tropical red loams of a lateritic character have been recognised in the upland regions, but erosion and human interference have been active in suppressing maturity of development.

There is a group of soils occupying the central plains which is characterised by an horizon of calcium carbonate accumulation. They appear to be poorly drained and may possibly prove to be assignable to the meadow soils group. Shaw, however, appears to regard them as belonging to Marbut's group of pedocals.

Indo-China. Tropical red loams and lateritic red earths occur in Indo-China.

AFRICA.

The soils of Africa lie almost entirely within the tropics and sub-tropics. Analogues with the soils of cool temperate regions are only to be expected in high elevations.

H. L. Shantz and C. F. Marbut, in a study of the vegetation and soils of the continent, have collected a considerable amount of information as to the principal types of

soil and their distribution. The map, prepared by Marbut, may serve as a starting point, although it is certain that many details and, possibly, some of the principal features of it may need modification.

The northern part of Africa is predominantly arid, a large proportion being actually occupied by the Sahara, Libyan, and Egyptian deserts. The soils of the Mediterranean seaboard include pedocals of the chestnut-coloured earth group passing through grey-brown desert soils to desert soils proper. In the wetter regions of the Atlas range, soils of the brown earth type may occur. South of the desert region the humidity increases and there is a transition through chestnut-coloured earths and tshernosems to the humid tropical soils of Equatorial Africa. The African pedocals have been little studied except in the Sudan. The soils of the Sudan Gezira, which have been so thoroughly studied by A. F. Joseph and his collaborators at Khartum, appear to be grey-brown soils of the semi-desert, in which the profile has been considerably modified by circulation consequent on the formation of deep fissures during drought.

Pedocalic soils occur also in Kenya, Somaliland, Abyssinia, and Rhodesia. Marbut's classification of the Transvaal "black turf soils" as tshernosems has been controverted by Marchand, and it is possible that equally intensive local study might necessitate further modifications in the provisional grouping suggested.

Saline and alkaline soils occur in association with the alluvial soils of the Nile delta and with the pedocalic soils of the Sudan.

The more humid parts of the continent are mainly occupied by red soils which appear to show all gradations between non-friable red loams and laterites. The latter group appears to be best developed in Sierra Leone, Liberia, and Ivory Coast.

Marbut distinguishes a group of soils as the Natal loams. These appear to be essentially similar to the red loams of lower latitudes. The soils of the Karroo and the

Kalahari desert are considered to belong to the brown-desert group. The soils of the humid Cape peninsula and the south coastal regions are essentially brown earths. The present writer has observed definitely podsol profiles developed in soils derived from Table Mountain Sandstone near Cape Town, and it is possible that they occur elsewhere.

The influence of parent material in modifying soil types in S. Africa has already been noticed. A further instance from the writer's own observation may be adduced from Natal, where igneous rocks and sands give rise to bright red soils, whilst, under the same climate, shales weather to brownish-grey soils. The essential factor may be, as in the case of the Transvaal soils, differences in drainage conditions.

Vlei soils are found in all parts of Africa, south of the Equator, and depend on local conditions of topography. Some examples have already been given in Chapter XI.

AMERICA.

United States. The great variety of climate, topography, and geology in the United States is reflected in its soils, which include representatives of nearly all the world groups with the exception of tundra and, possibly, laterite.

The most important fact to be noted is the boundary line between the pedocals and the pedalfers (leached soils), which runs approximately north and south through Minnesota, Eastern Nebraska, Mid-Kansas, Mid-Oklahoma, and Mid-Texas.

To the immediate east of the great soil boundary from Minnesota to Northern Texas, lie the so-called prairie soils, most typically developed in Missouri. These are considered by the American workers to be a sub-group of the podsols, whilst European workers have considered them to be either meadow soils or degraded tshernosems. Interspersed with the prairie soils are areas of meadow soils, as for example, in Iowa.

Podsols in varying stages of maturity occupy most of the northern portion of the pedalfer region, including

Eastern Minnesota, Wisconsin, Michigan, Northern New York, and the New England states.

The Middle-Eastern region is mainly occupied by brown earths, whilst the South-Eastern region from Virginia through the Carolinas, Georgia, Mississippi and Alabama, is mainly characterised by red soils with sub-tropical affinities. There are considerable areas of yellow soils in New Jersey and Maryland. These may form a transition between the brown earths and the sub-tropical red soils of the Southern States.

South of the 61°F isotherm in Alabama, and presumably also in Florida, the weathering is considered by L. D. Baver and G. D. Scarseth to be of the lateritic type, but it is doubtful if laterite profiles, such as those of the tropics, characterised by superficial incrustations, are present. The relative enrichment of the upper horizons in sesquioxides by removal of silicic acid is, however, well established.

Large stretches of barren sands also occur in the South-East. These, presumably, exhibit podsolisation. A considerable area of Florida is occupied by sub-tropical swamp soils.

West of the great boundary, increasing aridity is reflected in the successive belts of black earths and chestnut earths which pass in the northern region to grey desert soils with associated saline and alkaline soils. The latter groups attain their greatest development in Utah, Nevada, and Southern California. The southern desert soils are of the red and reddish-brown type.

Both in the mountain region and on the Pacific slope, are great areas of immature mountain soils. Brown earths, and, possibly, pod sols occur in Oregon and Washington.

Canada. The soils of Eastern Canada adjacent to the United States are probably pod sols, passing to tundra in more northern latitudes.

The course of the great soil boundary in Canada is somewhat indeterminate, but probably turns westward. Black earths, prairie soils and even pod sols occur in Manitoba,

Saskatchewan, and Alberta. Certain soils described as black prairie appear to belong to the meadow soil group. Saline soils occur in Alberta.

The soils of British Columbia are largely immature, but probably include brown earths and podsols.

All the groups mentioned pass northwards into the tundra area.

Central America and West Indies. The most detailed study of the soils of these regions is the survey of the soils of Cuba by Bennett and Allison, to which reference has already been made at numerous points in this book. Whilst the general tendency, with the prevalence of humid conditions, is towards the production of a sesquioxidic type of soil, all gradations can be observed from the well-known Nipe clay, which consists almost entirely of sesquioxides, to clays such as those of the Truffin and Bernal series in which a siliceous type of weathering complex is present.

Limestones give rise to two groups of soil. On the one hand are the mature leached red soils, exemplified by the Matanzas clay, which are essentially red earths; whilst the unleached limestone soils are represented by the dark coloured plastic Bayamo clay.

Among the soils of Western Cuba are soils which closely resemble certain series encountered in the South-Eastern United States. They are found in coastal plain areas developed from out-wash material.

There are large areas of poorly drained soils which have affinities with meadow and vleis soils. In certain cases layers of ferruginous concretions or even hardpan (mocarrero) are developed. Soils with layers of carbonate accumulation occur, but it is doubtful if they are to be grouped with the pedocals.

In Cuba, as in other parts of the West Indies, there are considerable areas derived from coral limestone. Some islands, e.g., Barbados, consist almost entirely of coral limestone soils. The contrast between red and black limestone soils has been discussed in Chapter XIV.

Many of the soils of the West Indies are very immature, whilst, in other cases, it seems likely that considerable erosion has taken place. There are extensive areas of clay soil in the Naparima district of Trinidad in which it would appear that the cultivated soil is only slightly modified from the parent geological material, which consists of marl or clay. Where maturity can be demonstrated, it is evident that the weathering is of a lateritic type. The most characteristic examples are seen in the weathering products of igneous rocks.

The soils of Central America, developed under humid tropical conditions, are generally similar to those of the West Indies, so far as can be judged. Pedocalic soils occur in Mexico, but have been little studied.

South America. The information at present available is too scanty to permit of anything more than broad generalisations to be made, but it is certain that representatives of all the great soil groups are present. Tropical red loams, red earths and laterites occur in Brazil and the adjoining countries east of the Andes. Soils of the pedocalic class (black earths, etc.) are found in the Pampas of Argentina, Uruguay, and Paraguay. Desert, saline, and alkaline soils occur in the coastal regions west of the Andes. Podsollic soils in varying stages of maturity occupy the extreme south.

AUSTRALIA.

As may be expected from the great range of latitude and the wide variations in humidity, the soils of Australia include representatives of all the important world groups with the exception of tundra.

A large part of Central and Western Australia is occupied by semi-desert and desert soils. Towards the east and north, more humid conditions obtain. The transition to the east is through grey and brown semi-desert soils and black earths, to red-brown soils, red loams, and podsollic soils which may be similar to the brown earths of the northern hemisphere. Red loams and laterites occur in the northern

regions. Laterites are strongly developed in South-Western Australia, where there are also considerable areas of salt-bearing soils known as Mallee soils. These occur also in South Australia and Victoria.

Care should be exercised in assigning Australian soils to the groups recognised in other parts of the world. Certain of the groups described, for example the Mallee soils, may, after further investigation, prove to be peculiar to Australia.

EAST INDIES.

Our information concerning the soils of the East Indies relates principally to Java and Sumatra; but much of what may be said of the soils of these islands probably holds, *mutatis mutandis*, for the other islands.

Although Java and Sumatra are entirely within the tropics, generally under a monsoon climate, there is, nevertheless, a marked vertical zonation of soils from the tropical lowland soils to veritable alpine soils. Both sedimentary and igneous rocks occur, and there is, therefore, the possibility of great complexity in the resultant soils.

Most of the soils appear to be relatively juvenile, either through erosion, or else on account of their development from recent volcanic material. The effect of parent material is very marked, but the general tendency is towards the formation of red loams and red earths.

In Middle and Eastern Java, under a relatively dry climate, there are large areas of dark-coloured soils, some derived from igneous rocks, and others from marls and limestones, which appear to have resemblances to the black cotton soils of India.

Alluvial and swamp soils are of common occurrence in the coastal lowlands and may, in some areas, be of a peaty character. Here, as in the uplands, the character of the parent material is frequently of the highest significance in determining the properties of soils, particularly their behaviour under cultivation.

BIBLIOGRAPHY.*

1. STREMMER, H., and AARNIO, B. (Editors), *Mémoires sur la nomenclature et classification des sols*. IV-ième Commission intern. pédol., Helsingfors, 1924.
2. MURGOCI, G. (Editor), *Etat de l'étude de la cartographie du sol dans divers pays*. V-ième Commission intern. pédol., Bucharest, 1924.
3. International Society of Soil Sci., *Comm. IV. General Soil Map of Europe, 1927*; with memoir by H. STREMMER. English edition by W. G. OGG, Macaulay Institute of Soil Research, Aberdeen, 1929.
4. GILCHRIST, D. A., and LUXMOORE, C. M., *Soils of Dorset*. Univ. Coll., Reading, and Dorset C.C., 1907.
5. FOREMAN, F. W., "Soils of Cambridge." *J. Agric. Sci.*, 1907, **2**, pp. 167-182.
6. HALL, A. D., and RUSSELL, E. J., *Agriculture and soils of Kent, Surrey, and Sussex*. Board of Agriculture, London, 1911.
7. NEWMAN, L. F., "Soils and agriculture of Norfolk." *Trans. Norfolk and Norwich Nat. Soc.*, 1912, **9**, pp. 349-393.
8. GOODWIN, W., *Soils of Nottinghamshire*. Notts. C.C., Nottingham, n.d.
9. ROBINSON, G. W., *Survey of the soils and agriculture of Shropshire*. Salop C.C., Shrewsbury, 1912.
10. RIGG, T., "Soils and crops of the market-garden district of Biggleswade (Beds.)." *J. Agric. Sci.*, 1916, **7**, pp. 385-431.
11. BRADE-BIRKS, S. G., and FURNEAUX, B. S., "Soil profiles in Kent." *J. South-Eastern Agric. Coll.*, 1928, pp. 224-234.
12. DAVIES, W. M., "Soils of Shropshire." *Agric. Progress*, 1928, **5**, pp. 50-56.
13. TEMPLE, M. S., "Survey of the soils of Buckinghamshire." *Univ. Reading*, 1929, *Bull.* 38.
14. SMITH, A. M., "The soils of Lancashire and Cheshire." A preliminary study. *J. Ministry of Agric.*, 1929, pp. 326-335.
15. PIZER, N. H., "Survey of the soils of Berkshire." *Univ. Reading*, 1931, *Bull.* 39.

*The references given here do not include all published works on the distribution of soils, nor, indeed, all the works consulted for the purpose of the above account. A fairly complete list is given, however, in the case of British soils since, in all probability, these will be of principal interest to the majority of readers.

Apart from the specific references given, a considerable body of information relating to soil cartography in many lands will be found in the volume of memoirs on soil cartography collected in 1924 by the late Professor G. Murgoci, of Bucharest, whose generous enthusiasm for the cartography of soils mainly inspired the co-operative efforts of the past decade, and, in particular, the production of the first Soil Map of Europe. Further information will be found in the volume of memoirs on soil nomenclature and classification collected by Professors H. Stremme and B. Aarnio in 1924. Finally, a considerable amount of information relative to the Russian Empire is to be found in the late Professor K. D. Glinka's "Typen der Bodenbildung."

16. NEWLANDS, G., "Scottish soil types with special reference to N.E. Scotland." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 194-199.
17. OGG, W. G., "Scottish soils in relation to climate and vegetation." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. I, pp. 301-309.
18. BERRY, R. A., LOUDEN, C., and MELVILLE, E. M., "Soils and agriculture of N. Ayrshire; in Geology of N. Ayrshire." *Memoirs of Geol. Survey*, 1929.
19. ROBINSON, G. W., "Studies on the Palæozoic soils of N. Wales." *J. Agric. Sci.*, 1917, **8**, pp. 338-384.
20. ROBINSON, G. W., and HILL, C. F., "Further studies on the soils of N. Wales." *J. Agric. Sci.*, 1919, **9**, pp. 259-282.
21. ROBINSON, G. W., JONES, J. O., and HUGHES, D. O., "Soil survey of Wales. Progress report, 1925-27." *Welsh J. Agric.*, 1928, **4**, pp. 303-321.
22. ROBINSON, G. W., HUGHES, D. O., and JONES, BRYNMOR, "Soil survey of Wales. Progress report, 1927-9." *Welsh J. Agric.*, 1930, **6**, pp. 249-265.
23. JONES, G. H. G., "Preliminary soil survey of the Creuddyn peninsula." *Welsh J. Agric.*, 1927, **3**, pp. 154-164.
24. ROBINSON, G. W., "Development of the soil profile in N. Wales as illustrated by the character of the clay fraction." *J. Agric. Sci.*, 1930, **20**, pp. 618-639.
25. BJÖRLYKKE, K. O., "Soil types and soil profiles in Norway." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 223-269.
26. BJÖRLYKKE, K. O., "Om Norges Jordsmonn" (with English summary). *Norsk geologisk tidsskrift*, 1931, **12**, pp. 89-116.
27. GLINKA, K. D., *Die Typen der Bodenbildung*. Borntraeger, Berlin, 1914.
28. TAMM, O., "Die Klimatischen Bodenregionen in Schweden." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. I, pp. 269-285.
29. ANDERSEN, J., "Classification du sol de Danemark"; in *Mémoires sur la nomenclature et classification des sols*, Helsingfors, 1924.
30. AARNIO, B., "Chemical and physical properties of Finnish soils." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 507-523.
31. FROSTERUS, B., "Die Klassifikation der Böden und Bodenarten Finnlands"; in *Mémoires sur la nomenclature et classification des sols*, Helsingfors, 1924.
32. STREMMER, H., "Die Böden Deutschland" in BLANCK's *Handbuch der Bodenlehre*. Band V, pp. 271-429.
33. SMOLIK, L. F., "Principal soil profiles of Czechoslovakia." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 323-327.
34. JENNY, H., "Die alpinen Böden," *Denkschr. schweiz. naturforsch. Gesell.*, 1926, **63**, pp. 297-340.
35. JENNY, H., "Soil investigations in the Swiss Alps." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 421-428.
36. MIECZYNSKI, T., et. al. *Materjaly do poznania gleb Polskich* (German Summary), Cracow, 1925.
37. BALLENEGGER, R., "Sur la composition chimique des sols de la Hongrie." *Actes IV-ième Conf. Intern. Pédol.*, 1926, Vol. III, pp. 410-433.

38. DE SIGMOND, A. A. J., *Hungarian alkali soils and methods of their reclamation*. Univ. of California, Berkeley, 1927.
39. TREITZ, P., "Les regions naturelles du sol de la Hongrie historique et les types des sols principaux, en relation avec la classification générale." *Actes IV-ième Conf. Intern. Pédol.*, 1926, Vol. III, pp. 434-439.
40. AGAFONOFF, V., "Les types des sols de France." *Soil Res.*, 1928, **1**, pp. 67-91.
41. HISSINK, D. J., "Versuch eines Nomenklatur und Klassifikation der niederländischen Böden, nebst Bemerkungen der Methodik." *Actes IV-ième Conf. Intern. Pédol.*, 1926, Vol. III, pp. 474-479.
42. DEL VILLAR, E., "Esquisse préalable de la distribution des types de sols en Espagne." *Proc. 1st Int. Congr. Soil Sci.*, Vol. I, pp. 310-317.
43. DEL VILLAR, E., "Les sols méditerranéens étudiés en Espagne." *Proc. 1st Int. Soc. Soil Sci.*, 1930, **5**, pp. 192-197.
44. BOUYOUCOS, G., "A study of the fertility of the soils of Greece." *Soil Sci.*, 1922, **13**, pp. 63-79.
45. SAHASRABUDDHE, D. L., "Soils of the Bombay Presidency." *Dept. Agr., Bombay*, 1929, *Bull.* 160.
46. SHAW, C. F., "Soils of China." *Geol. Survey of China*, 1930, *Soils Bull.* 1.
47. AGAFONOFF, V., "Sur quelques sols rouges et Bienhoa de l'Indochine." *Soil Res.*, 1930, **2**, pp. 184-195.
48. SHANTZ, H. L., and MARBUT, C. F., *The vegetation and soils of Africa*. Amer. Geogr. Soc., New York, 1923.
49. HUME, W. F., "Character of the soils of Egypt"; in *Mémoires sur la nomenclature et classification des sols*. Helsingfors, 1924.
50. MARCHAND, B. DE C., "Origin of the black turf soils of the Transvaal." *S. Agr. J. Sci.*, 1924, **21**, pp. 162-181.
51. MAUFE, H. B., "On the formation of red soil and of black vlei soil from dolerite at Salisbury, Rhodesia." *S. Afr. J. Sci.*, 1928, **25**, pp. 156-167.
52. GREENE, H., "Soil profile in the eastern Gezira." *J. Agric. Sci.*, 1928, **18**, pp. 918-930.
53. GRACIE, D. S., "A preliminary survey of some of the soils of Kenya." *Dept. Agr. Kenya*, 1930, *Bull.* 1.
54. MARBUT, C. F., "A scheme for soil classification." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 1-31.
55. VILENSKY, D. G., "Principal features of distribution of soils and vegetation in the United States." *Soil Res.*, 1928, **1**, pp. 108-137.
56. SHAW, C. F., "The basis of classification and key to the soils of California." *Proc. 1st Int. Congr. Soil Sci.*, 1928, pp. 65-103.
57. NORTON, E. A., "Profiles of soils in southern Illinois." *Proc. 1st Int. Congr. Soil Sci.*, 1928, pp. 283-290.
58. KELLOGG, C. E., "Preliminary study of the profiles of the principal soil types of Wisconsin." *Wis. Geol. Nat. Hist. Survey*, 1930, *Bull.* 77.

59. JOEL, A. H., "Predominant Saskatchewan soil profiles correlated with soil development factors in northern latitudes." *Proc. 1st Int. Congr. Soil Sci.*, 1928, pp. 200-222.
60. WYATT, F. A., and NEWTON, J. D., "Alberta soil profiles." *Proc. 1st Int. Congr. Soil Sci.*, 1928, pp. 358-366.
61. MATTHAEI, A., "Über die Genesis der Böden von Chili." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 536-539.
62. MATTHAEI, A., "Einfluss von Oberflächengestaltung, Klima, und Vegetation auf die Verbreitung der Bodentypen in Chile." *Soil Res.*, 1930, **2**, pp. 57-67.
63. BENNETT, H. H., "Some comparisons of the properties of humid-tropical and humid-temperate American soils; with special reference to indicated relations between chemical composition and physical properties." *Soil Sci.*, 1926, **21**, pp. 349-374.
64. BENNETT, H. H., and ALLISON, R. V., *The soils of Cuba*, Washington, 1928.
65. FIPPIN, E. O., "Soils of the Republic of Haiti." *Proc. 1st Int. Congr. Soil Sci.*, 1928, Vol. IV, pp. 270-275.
66. HARDY, F., AKHURST, C. G., and GRIFFITH, G., "The Cacao soils of Tobago." *Supplement to Tropical Agriculture*, 1931, pp. 1-22.
67. PRESCOTT, J. A., "The soils of Australia in relation to vegetation and climate." *Council for Ind. Sci. Res., Australia*, 1931, Bull. 52.
68. BRYAN, W. H., and HINES, H. J. G., "Factors in the development of soil profiles in Southern Queensland." *Soil Res.*, 1931, **2**, pp. 264-276.
69. MOHR, E. C. J., *Tropical soil forming processes and the development of tropical soils, with special reference to Java and Sumatra*. English translation by R. L. Pendleton. University of the Philippines, 1930.
70. SENSTIUS, M. W., "Agrogeological studies in the tropics." *Soil Research*, 1930, **2**, pp. 10-56.

CHAPTER XVII.

SOIL SURVEYS.

AIMS AND METHODS.

WHILST the aims of soil surveys are often mainly of an utilitarian character, work of this type is of great importance in elucidating the regional relationships of soils and the influence of climate and topography on soil genesis.

Soil surveys have been in progress for many years in different countries, but the organisation of soil survey work has been developed more thoroughly than elsewhere in the United States of America in the form of a national soil survey under the Federal Government. In other countries, surveys have been made by workers at provincial institutions, or soil survey work has been included in the activities of the Geological Survey, as in Germany. In South Africa, there is a soil survey organisation specially devoted to irrigated and irrigable lands.

It is natural that, in different countries, there should be considerable variation in the aims and methods of soil surveys. To some extent, the degree of complication encountered in the distribution of soils will affect the methods chosen for their survey. The soil profile, the natural unit of study, is often so variable in its character, reflecting, as it does, changes in parent material, climate, and topography, that the execution of a complete record by mapping all relevant data may be a very slow procedure. On the other hand, there are areas even in Britain, where relatively simple geology and topography under a uniform climate result in extensive tracts of soil which, if not actually constant in

character, exhibit variations which can be easily related to topographical conditions.

We have already discussed the broad world classification of soils. In the present state of our knowledge, this classification is still imperfect owing to the insufficiency of existing information as to the soils of large areas of the globe. And, therefore, whilst it is possible, in countries whose soils have been minutely studied, to make direct use of some of the distinctions of a world classification for the purpose of primary study, in other countries the position of their soils in a world system is as yet so uncertain that it is not always possible to assign them to known groups. It is, indeed, certain that fuller knowledge of hitherto unexplored soil regions will add materially to our list of soil groups.

Whilst the survey of the soils of a new country should be carried out with careful attention to relationships with world groups, the first requisite is to obtain such a classification that the soils of the particular country will be thrown into the clearest possible relationship to each other. A valid classification of the soils of a given country is a necessary preliminary to their assignment to positions in a world system.

The scale of actual mapping varies considerably. In intensive surveys of highly cultivated areas, it may be necessary to map on as large a scale as 25 in. to the mile. For ordinary agricultural purposes in this country, a scale of 6 in. to the mile is more convenient. Smaller scales may be more useful for the rapid reconnaissance of large areas. The choice of a scale will, of course, be governed to some extent by the base maps available.

CHARACTER OF DATA COLLECTED.

The character of the data collected in a soil survey will vary according to the amount of information already collected by other organisations. In Great Britain, much information relevant to the soil is already recorded on the Ordnance Survey maps. Fairly detailed information as to

the geology is generally available, although in many cases the surface geology, which is of principal importance to the surveyor, is rather imperfectly described. In less settled countries, the amount of information available from other sources may be very scanty, and the surveyor is responsible for the collection of all data.

From whatever sources they are obtained, the following data appear to be necessary for the characterisation of soil conditions.

- A. Surface features.
- B. Soil water conditions.
- C. Texture.
- D. Colour.
- E. Depth and succession of soil horizons down to parent material.
- F. Character of parent material.
- G. Nature of cropping or vegetation.

A. Surface Features. Much of the information under this heading is already shown, so far as Great Britain is concerned, on the Ordnance maps used as base sheets. Where, however, contour lines are at wide intervals in gently undulating country, important surface features may be unrecorded on the base maps and should be shown on the soil map. It is convenient to distinguish a small number of categories for record. In Great Britain, for example, the distinctions, flat, rolling, steep, very steep, and broken or irregular are noted by appropriate symbols. Arrows are inscribed on the maps to show directions of slope. Special symbols may be used to denote broken country, knolls, and depressions.

B. Soil Water Conditions. Both from the standpoint of soil classification and of practical utility, distinctions under this head are of the greatest importance. It is convenient in the first place to distinguish soils of free drainage from soils with impeded drainage. The former may be subdivided into soils which are satisfactory and soils liable to drought. Soils with impeded drainage may be divided into

soils with positional water-tables as in the valley bottoms or coastal regions, and soils of elevated regions having impervious strata. Varying degrees of wetness may be distinguished and, in addition, it should be noted whether wetness is continuous or intermittent.

It sometimes happens that excessive wetness may alternate with excessive dryness in the same soil. Changes in level of the water-table in a coarse sand may lead to such alternations. Further, in heavy clay soils, winter wetness may have the effect of restricting root development to the surface horizon. When drought occurs, the moisture within root range is quickly exhausted and capillary rise is too slow to repair the losses by evaporation and transpiration. These circumstances are generally reflected in the character of grass herbage. Indeed, in a grass country, herbage is often the surest key to the nature of the soil.

C. Texture. For some purposes, distinctions in texture are of great importance. For ordinary soil surveys, however, it is not practicable to have more textural grades than can be distinguished in the field by personal judgement. In Great Britain, the following texture grades are recognised:—

Light Sand	...	...	...	...	Sa
Heavier Sand (or Loamy Sand)	...	...	...	...	Sb
Light Loam	...	...	...	...	La
Medium Loam	...	...	...	...	Lm
Heavy Loam	...	...	...	...	Lb
Clay	...	...	...	...	C
Coarse Silt	...	...	...	...	Za
Fine Silt	...	...	...	...	Zb

Intermediate grades may also be distinguished, such as sandy clay (Cs) and clayey sand (Sc).

In the United States, a much larger number of textural grades is recognised.

The judgement as to texture obtained in the field is checked from time to time by actual mechanical analyses in the laboratory. With experienced surveyors, the correspondence is fairly close.

D. Colour. Colour is perhaps the most difficult of all soil properties to describe accurately. Up to the present, no generally accepted scheme of colour description has been agreed upon. Such a scheme would be of considerable value, for there is, perhaps, no other soil property which is of such value for diagnostic purposes in the field. At present, the recognition and recording of colour distinctions is generally subjective and personal.

The most important distinctions to be noted in soil surveying are between grey colours, on the one hand, and red and brown colours, on the other. Grey colours are found in arid soils, in the bleached layers of podsol soils, in soils with impeded drainage, and, in Britain and Northern France, in immature or eroded chalk soils. Brown and red colours are found in tropical soils and also in soils of temperate climates formed with complete leaching in the absence of acid humus. Red colours may be conferred by the parent rock as in the soils of the Trias and Old Red Sandstone in England. Experienced surveyors can frequently detect the presence of admixtures of drift from the colours of soils.

E. Profile Character. The unit of soil study is the complete succession of horizons down to the parent material from which they have been differentiated by the pedogenic processes. The soil survey should, therefore, include descriptions of all types of soil profile encountered. These are conveniently recorded in notebooks with corresponding references on the map. The data should include information as to depth, texture, colour, soil reaction, and moisture conditions down to and including the parent material.

In routine soil surveys, profile characters are usually ascertained by boring with an auger. The actual depth of exploration will vary somewhat, but should reach to the full extent of the soil horizons. It is not generally necessary in England and Wales to go below four feet in order to reach the parent material, but it is desirable, at intervals, to carry the examination to greater depths for confirmation. In addition, in default of trustworthy sections, holes or trenches

should be dug in typical places in order to be able to observe the natural structure of the profile. Such holes can be conveniently used for sampling the several horizons.

F. Parent Material. The importance of parent material in the recorded data of a soil survey is two-fold. In the first place, although the parent material does not strictly belong to the soil horizons, yet it frequently has an important influence on the moisture relationships within the soil profile as, for example, in the soils of the Chalk in southern England. In immature soils, the profile may consist entirely of slightly modified parent material without true soil horizons. This is the case in recent alluvial soils such as those of recent reclamations in Holland. Secondly, parent material is an important aid to classification. Although the basis of classification is the actual soil profile, it is convenient to be able to associate a particular series of soils with a given parent material. The parent material is not so much a part of the definition of a soil class as an indication of where soils of that class may be expected to occur.

Where similar conditions of profile development occur, it is found that rocks of similar lithological character, even though stratigraphically distinct, give rise to similar soils. The Powys soils of Wales are associated with sedimentary rocks from the Cambrian to the Silurian. Further study may reveal other instances in which rocks of different ages may give rise to similar soils. The criterion is always the soil profile.

G. Natural Vegetation and Cropping. In a new country, a record of the natural vegetation is of the utmost importance, and may provide important clues to the distribution of soils. In difficult country, aerial photography has proved to be a convenient method of mapping vegetative cover if the aerial maps are compared with accurate ground surveys of selected areas.

In settled countries, the record of cropping at the time of survey is only of fugitive significance, but may be important in connexion with advisory work. In Great Britain,

data on grass herbage are of considerable value both from the practical standpoint and also as an aid to the recognition of certain soil characters, more particularly the base status and the moisture relationships.

Where possible, notes on weed and hedgerow flora should be made. Tree growth, also, can often be correlated with soil characters.

In the uncultivated lands of Wales, an exceedingly close correspondence is found between soil characters and vegetation type. By a view of the vegetation alone, it is frequently possible to obtain a fairly correct idea of the distribution of the different classes of soil.

EFFECT OF SUPERFICIAL GEOLOGY AND TOPOGRAPHY.

The problems confronting the soil surveyor will vary with the region to be surveyed. Whilst, in some areas, soil differences are easily seen, and the different classes recognised can be readily correlated with geology and topography, in other areas, the changes in soil character are so frequent that a trustworthy map can only be produced by the most minute examination. Complications, in such cases, generally reflect the intricacies of superficial geology. This obtains particularly over large areas of England and Wales, where the main task confronting the surveyor is the perception and delineation of glacial, colluvial, lacustrine and alluvial deposits. Considerable help can often be obtained by an intelligent and imaginative interpretation of topography; yet there are also areas, such as certain alluvial and lacustrine flats, where the facts can only be ascertained by closely ranged auger examinations. In such cases, it may be impracticable to insert all the existing detail, but a close survey of specimen areas may serve to indicate the kind of variation to be expected over the area as a whole.

CLASSIFICATION INTO SERIES.

The arrangement into categories is based on the profile characteristics. Soils with similar profiles derived from

similar material under similar conditions of development are conveniently grouped together as *series*. A series is further sub-divided according to textural variations. In the United States of America, where this system is used in the Soil Survey, series are named after the localities where they are first studied or where they attain considerable development. The textural distinctions are shown in the *types*. Thus, the Leonardtown series include such types as the Leonardtown medium loam, the Leonardtown fine sandy loam, and so on.

The series system with some adjustments has been adopted in this country. In the soil survey of Wales, it has been found convenient to make a preliminary classification into *suites*, i.e., soils derived from the same or similar parent material. The qualification "or similar" has been introduced because it has been found that parent materials of different geological age, but of generally similar lithological character, give rise to the same type of soil profile under similar conditions of formation. The soils of a suite are divided into series according to the mode of development, as reflected in the profile. For example, the Powys suite comprises soils derived from Cambrian, Ordovician, and Silurian non-calcareous sedimentary rocks, excluding hard crystalline grits. The following series have, up to the present, been recognised:—

- (1) Powys Series. Well drained soils, sedentary on the parent rock.
- (2) Penrhyn Series. Well drained soils derived from glacial drift or colluvial material.
- (3) Cegin Series. Soils with impeded drainage derived from glacial drift or colluvial material.
- (4) Conway Series. Alluvial soils.

Further series will doubtless be recognised; for example, the soils showing podsolised or eroded profiles in the uplands. Each series gives types according to texture. Whilst, theoretically, a series might include a complete textural range, the variation is comparatively small in practice.

In the early stages of a survey, it is generally desirable

to avoid too much restriction of the number of series. It is safest to assume, in the first instance, that different parent materials, even though lithologically similar, give rise to distinct soils. When, however, further study shows that the soils are indistinguishable, the series may be combined.

For example, in the author's experience, there is no essential difference between the soils from similar Cambrian, Ordovician, and Silurian sediments, and they are therefore grouped together. If it could be demonstrated that soils derived from Devonian or Carboniferous shales are essentially similar to these, then they also could be included in the larger group. The criterion is always the soil profile.

USE OF LABORATORY DATA IN SURVEY WORK.

The first estimate of the soils of an area is, of course, made in the field, and in the opinion of some surveyors, soils should be mapped on field characteristics. Yet the identity of a given soil profile must rest ultimately on some securer basis than qualitative and personal judgement. It is therefore desirable that representative profiles should be submitted to careful laboratory examination and that the series and type characters should be defined in a quantitative manner.

A complete definition of a series should give data on the following points:—

(1) Mechanical composition of the principal types. Whilst a certain textural range within the series is admissible, it is often the case that the mechanical composition curves show a general similarity which suggests their relationship. Mechanical analysis should always be used as a check on field estimates of texture.

(2) Organic carbon and nitrogen content. These figures will show some variation, yet the vertical variation of the organic carbon content within the profile may prove to be characteristic of a series. For example, soils with impeded drainage frequently show a sharp decrease in organic carbon content at the horizon of impedance.

(3) The composition of the clay fraction. The changes in the composition of the clay fraction, as reflected in the silica-sesquioxide and alumina-ferric oxide ratios afford a valuable indication of the nature and extent of the processes of eluviation which have taken place. Further, it has been repeatedly shown that the properties of soils are deeply influenced by the nature of their clay fractions. Highly siliceous clays exhibit strongly colloidal properties, whilst clays rich in sesquioxides are markedly less colloidal. In the writer's opinion, the variations in the composition of the clay fraction, particularly the silica and sesquioxides, afford the most valuable information of all the analytical determinations made.

(4) The base status. The variations in the base status as shown by the content of exchangeable bases or by the pH, are of great diagnostic value. In cultivated soils, however, liming may produce modifications, of which, however, the surveyor should be aware.

(5) The nature and distribution of soluble salts. Whilst these do not occur to any marked extent in British soils, there are countries such as the Sudan, S. Africa, and Australia, where information on this point is of the greatest value in characterising and classifying soils. Determination of exchangeable sodium may give valuable information.

(6) Mineralogical composition. A knowledge of the minerals present in a soil is of great diagnostic value, provided the mineralogical composition of the parent material is also known. R. Hart¹, in Scotland, has found the minerals of great assistance in identifying glacial drifts. Further, under tropical conditions, a knowledge of mineralogical composition may be useful in estimating the extent to which the soil has arrived at the end-point of its weathering. Unfortunately, available methods of analysis, in addition to being very laborious, do not succeed in giving a complete picture of the mineralogical composition of the soil, for they are not applicable to particles of finer grade than fine sand. It is possible that X-ray methods may prove of assistance

in defining soils, but this line of attack is as yet in its early stages.

(7) Physical properties. It would be desirable to be able to express quantitatively those physical properties of a soil which affect its water-air relationships. Data showing the variation in pore space throughout a profile would be exceedingly valuable as a measure of the degree of compaction. Unfortunately, the physical analysis of soils is at present in a less advanced stage of development than the mechanical and chemical analysis.

We have already alluded to the importance of obtaining a satisfactory scale for the expression of soil colours. When such a scale is available it would be desirable that colours should be recorded both for wet and dry soil, and after removal of humus by peroxide treatment.

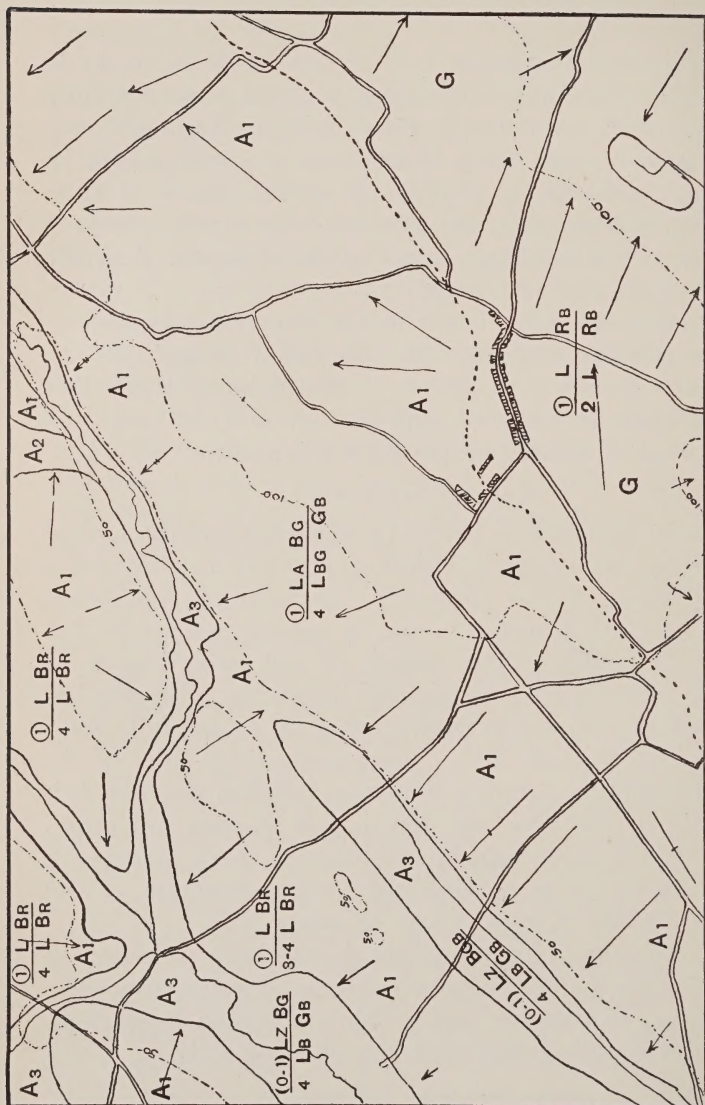
SOIL MONOLITHS.

In connexion with soil surveys in Russia and the United States, it has been found helpful to preserve records of the principal kinds of soil profile in the form of *soil monoliths*. These are actual columns of natural soil to the full depth of the profile. The taking of soil monoliths is an operation demanding considerable ingenuity. Essentially, the procedure consists in excavating a pit to the full depth of the soil profile and, after cutting away three sides of the proposed monolith, fitting a stout box. When the monolith fits snugly in the box, the fourth side is then cut away. The cut face is then chipped to obliterate spade marks and to throw up as far as possible the natural structure. The greatest difficulty is experienced in stony soils.

As a substitute for monoliths it has been proposed to apply a flexible material covered with an adhesive substance, such as glue, to a profile face. When adhesion is attained, the material is pulled away with the adhering soil.

PREPARATION OF MAPS.

In soil survey work in Great Britain, the Ordnance Survey map on the scale of six inches to the mile proves the



Surveyed by D.O. Hughes, J.O. Jones and Brynmor Jones.

FIG. 11.—Portion of six-inch soil map of Anglesey, XXII N.E.
(From "Journal of the Ministry of Agriculture"—by permission.)

Series. A1. Gaerwen loam to light loam; deep, well drained soils derived from metamorphic schists. A2. Gesall heavy loam; soils derived from schistose material with impeded drainage. A3. Braint silty loam; alluvial soils from schistose material. G. Gower loam; rather shallow, well drained soils derived from Carboniferous Limestone. The arrows point down slopes.

most convenient base map for use in the field. On it should be recorded all data which can be conveniently inserted without undue crowding. In complicated areas, however, it may be desirable to insert key-letters or numbers referring to descriptions made in a field note book. Boundaries between soil series should be drawn in the field.

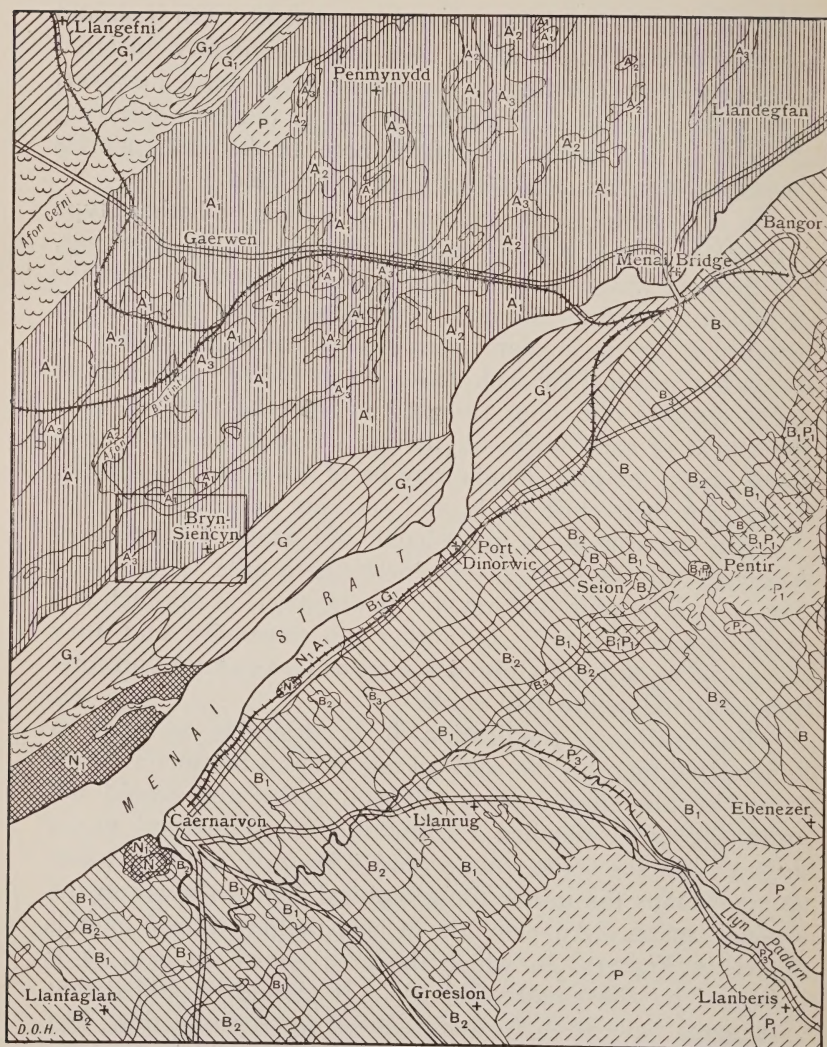
It is, perhaps, a matter for debate whether type boundaries should always be inserted. Generally speaking, the textural range within a soil series is not wide, and the delineation of fine textural distinctions may involve an amount of labour incommensurate with the value of the information given. In such cases it should be sufficient to insert textural data (La, Lb, etc.) at frequent intervals over the extent of each series on the map. In intensively cultivated areas, textural details may be of sufficient importance to warrant their close study and accurate delineation.

An example of a completed soil map of a part of Anglesey on the scale of six inches to a mile is shown in *Fig. 11*. The reproduction, for publication, of six-inch soil maps would generally be rather too expensive and they should be regarded mainly as the repositories of the information gathered in the field work.

For ordinary publication it is desirable to make a generalised map on the scale of one mile to the inch. Such a map would simply show the soil series and would not ordinarily carry detailed information as to textural variations within the series; nor would it be practicable to show on such maps the minor topographical details shown on the six-inch maps. An example of such a generalised map is shown in *Fig. 12*, in which the area shown in *Fig. 11* is indicated by the small rectangle. It should be added that both maps have been further reduced for reproduction; but the difference in the amount of detail shown will be evident.

BIBLIOGRAPHY.

- ¹ HART, R., "Studies in the Geology and Mineralogy of soils." I.—A detailed study of a region characterised by diverse rocks and partly covered by glacial drift. *J. Agric. Sci.*, 1929, **19**, pp. 90-105.



Surveyed by D.O. Hughes, J.O. Jones, Brynmor Jones, and W.G.D. Walters.

Emery Walker Ltd. sc.

0 1 2 3 4 Miles

- 
ANGLESEY SUITE: A₁, Gaerwen; A₂, Gesail; A₃, Braint.
- 
BANGOR SUITE: B, Bangor; B₁, Ebenezer; B₂, Sion; B₃, Glanadda
- 
GOWER SUITE: G, Gower; G₁, Pentraeth.
- 
NEATH SUITE: N, Neath; N₁, Ilston.
- 
POWYS SUITE: P, Powys; P₁, Penrhyn; P₂, Cegin; P₃, Conway.
- 
Estuarine Alluvium.

FIG. 12.—Soil series map of portions of Anglesey and Caernarvon.

CHAPTER XVIII

SOIL ANALYSIS.

AIMS OF SOIL ANALYSIS.*

IN the earlier work on the analysis of soils, the object in view was generally the discovery and estimation of manurial requirements. In the absence of clear conceptions of the nature of plant nutrition and of the constitution of the soil, it is not surprising that soil analysis failed to achieve the practical success expected by its first exponents. Even with the fuller knowledge now at our disposal, it must be admitted that most of the applications of soil analysis to practical ends are of an empirical character. Such, for example, are the methods for determining available plant nutrients by means of extraction with dilute acid solvents.

In considering methods of soil analysis, we may distinguish between *absolute* and *conventional* methods. In an absolute method, a determination is made of the total amount of a given constituent present in the soil, or of the amount falling in a definite category of the soil. The determination of the total nitrogen content of the soil is an absolute determination, as is also the determination of the ammoniacal nitrogen content. Similarly we may characterise the determinations of exchangeable bases as absolute determinations. Wherever it is possible to distinguish categories of constituents in the soil, the natural corollary is an absolute method of determination.

On the other hand, we have conventional methods. These are generally the outcome of practical requirements

*The purpose of the present chapter is to discuss the significance of soil analysis. The actual methods are given in the Appendix.

and are of slight significance apart from their practical utility. In the determination of available phosphoric acid by the citric acid extraction method, it is necessary to prescribe every detail of working; for any variation, as, for example, in the ratio of solvent to soil, or in the time of extraction, may lead to different figures being obtained. The phosphoric acid thus determined does not represent any category of soil phosphorus which can be defined in terms other than those of the method by which it is obtained. Methods such as these can only have a temporary value and must be superseded when fuller knowledge of the constitution of the soil enables us to distinguish and determine the different categories of soil constituents.

From the standpoint of the study of the soil as a pure science, the chief end of soil analysis is to characterise the material with which it deals. An ideal system of soil analysis should give a complete quantitative description of the soil, which will serve as a basis for comparison and classification. The need for such a system of analysis is abundantly evident when we consider the qualitative terms in which many of the most important groups of soils have been described. The task of soil classification cannot be considered complete until these subjective descriptions have been translated into definite quantitative data. Such data are as important as the physical and chemical data necessary for the description of chemical compounds.

MECHANICAL COMPOSITION.

The first information required about a soil is its mechanical composition. The practical man, confronted with a soil, will always wish to know whether it is a clay, a loam, or a sand, and the pedologist will seek the same information expressed in quantitative terms. Mechanical composition gives quantitative expression to a fundamental characteristic of soil, unaffected by the ordinary vicissitudes of cultivation and cropping. The results of mechanical analysis can be most readily interpreted when shown in the

form of a summation curve, for, as we have seen in Chapter II, different types of soil give characteristic curves. Thus, there are characteristic curves for highly weathered soils, for mechanically weathered soils, for alluvial soils, and for soils of mixed origin.

The most important single figure in mechanical analysis is the clay content. If the clay fraction has been appropriately defined, it enables us to state what proportion of a given soil consists of the reactive colloidal products of chemical weathering, as distinct from the unweathered material, which is relatively inert and acts simply as a framework or skeleton for the reactive constituents.

IGNITION LOSS.

The loss in weight suffered by an oven-dry soil on ignition was formerly used as a measure of the proportion of organic matter present. It was always realised, in the case of soils containing much clay or calcium carbonate, that the results were likely to be of little value, and even with lighter soils, small amounts of organic matter cannot be shown by ignition loss. Recently, however, this determination has acquired new importance, for B. A. Keen and J. R. H. Coutts¹ have shown that it does afford an approximate indication of the colloidality of the soil and is highly correlated with the " sticky point " value.

CALCIUM CARBONATE.

Calcium carbonate is usually determined and shown with the mechanical analysis of a soil. The presence of calcium carbonate, except where it occurs as comparatively coarse isolated fragments, indicates base-saturation and, in natural mature profiles, is characteristic of certain horizons of soils of the tshernosem and chestnut earth group. Calcium carbonate may also occur, under certain circumstances, in horizons affected with ground-water. Among immature soils, the rendzinas are characterised by the presence of free calcium carbonate as are, also, the related chalk

soils of South-East England and Northern France. In soils of the humid regions—the “ pedalfers ” of Marbut—calcium carbonate is only a fugitive constituent and does not occur in mature profiles.

ORGANIC MATTER.

Of equal importance with mechanical composition is a knowledge of the organic matter content of the soil. The greater proportion of soil organic matter consists of colloidal material which, with the clay, forms the colloidal complex. The organic matter is also the seat of microbiological changes in the soil. It is obvious, therefore, that a knowledge of the amount of this constituent is of the highest value in characterising a soil. Its direct determination is a matter of some difficulty, and it is more usual to obtain a measure of it from the determination of organic carbon, using a conventional factor (usually 1.724) to convert to organic matter. The organic carbon is determined by dry combustion or by wet combustion methods. In cultivated soils, it is sometimes necessary to make a correlation for elementary carbon in the form of coal or cinder.

The determination of the organic matter content is only the first approach to a characterisation of the organic portion of the soil. There are obvious differences between 5% of organic matter in a typical red loam, a black earth, and a podsol, respectively. It is, therefore, desirable to have methods for the fractionation of organic matter. Various methods have been proposed, such as the determination of humified organic matter by hydrogen peroxide, and the determination of pentosans, lignins, and celluloses; but, until the proposed methods have been investigated over a wider range of soils, the significance of the results yielded by them is not sufficiently evident for their adoption in routine analysis. It is probable, however, that investigations now in progress will eventually lead to the recognition of methods for the characterisation of organic matter which will be of the highest value in soil classification.

NITROGEN.

The determination of nitrogen is often made without any clear idea as to its significance. For purely practical purposes, it can rarely serve as a measure of the nitrogen status of the soil. As a measure of soil organic matter, it is highly untrustworthy because of the great variation in the nitrogen content of organic matter. In recent years, a considerable amount of work has been carried out on the carbon-nitrogen ratio of soils, and, whilst the significance of this ratio is not completely understood, some of the indications are sufficiently suggestive to render it highly desirable that it should be determined over as wide a range of soils as possible. It may yet prove a valuable aid to the definition of soil groups.

The significance of the ammoniacal nitrogen content of soils is not sufficiently evident to recommend its general determination. Except in acid organic soils, it is generally present in very small proportions. The nitrate nitrogen varies so greatly in amount in the same soil that it can scarcely be used as an aid to description.

TOTAL ANALYSIS.

In a perusal of the literature of the subject, the student will encounter numerous descriptions of soils and soil profiles in which the complete analysis of the total soil is given. Such analyses, involving the determination of silica, sesquioxides, titanium oxide, alkalies, and alkaline earths, represent a very considerable mass of effort, yet the value of the information to be derived from them is, in the writer's opinion, of an entirely lower order of magnitude, since they fail to distinguish between the unweathered minerals and the weathering complex. They may be compared with results of analyses of entire animals made by the pioneers in animal nutrition.

COMPOSITION OF THE CLAY FRACTION.

Whilst the total analysis of a soil may be, in itself, of comparatively little value for diagnosis, the analysis of the

clay fraction, presumed to be identifiable with the weathering complex, yields results of the highest value. For many purposes it is sufficient to determine silica, alumina, and ferric oxide. The results obtained are of significance not only from the standpoint of the genetic classification of soils, but also as a clue to soil properties.

It has been shown in the discussion of the pedogenic processes and in the description of the principal soil groups, that high silica-sesquioxide ratios are characteristic of soils developed under arid conditions, of soils with impeded drainage, of unleached limestone soils, of the A horizons of the podsoles and soloti soils, and of sub-aqueous sediments. On the other hand, low silica-sesquioxide ratios are associated with brown forest soils, tropical red soils, leached limestone soils, and the B horizons of podsoles.

Further, colloidal properties, including imbibitional moisture capacity, plasticity, and absorptive capacity, are highly correlated with the silica-sesquioxide ratio.

For these reasons, the writer would regard the silica-sesquioxide ratio of the clay fraction as, perhaps, the most important information obtained in the laboratory examination of soils.

DEGREE OF WEATHERING.

The writer has recently, in collaboration with Dr. M. Richardson, examined the possibility of estimating the degree of weathering of soils from the relationship between the total alumina and the alumina present in the clay fraction. Some preliminary results, shown on p. 50, indicate the possibilities of the method, which may prove of value for descriptive and diagnostic purposes.

EXCHANGEABLE BASES.

The properties of soils are, as we have already seen, so strongly affected by the nature and amount of the exchangeable bases present that any description of soil must include data for these constituents. For most soils of the humid

regions, about 80% of the reaction value of the total bases, exclusive of hydrogen, is represented by exchangeable calcium, and therefore it is generally sufficient to determine only this base. In saline and alkaline soils, however, exchangeable sodium forms a considerable if not principal proportion of the exchangeable bases, with or without free sodium salts. The determination of exchangeable bases in soils containing notable amounts of free calcium carbonate is generally unnecessary, except in saline and alkaline soils, but may be carried out on typical samples.

In most soils, calcium is the dominant exchangeable base, and a determination of exchangeable calcium gives a fair measure of the total exchangeable bases present. For more accurate work, it is often desirable to determine the other bases and also the reaction value of the sum of the exchangeable bases. Where this value is reached by summation of separate determinations, it is usually somewhat magnified by the inclusion of soluble salts, principally calcium sulphate and chloride. This is avoided in the acetic acid method of R. Williams².

REACTION.

The reaction of soils, generally expressed as pH, the negative logarithm of the hydrogen-ion concentration, has been the subject of a large amount of investigation. Before discussing its significance, it is necessary to direct attention to the somewhat conventional and arbitrary character of the results obtained. When we speak of the pH of a soil, we cannot achieve a precise significance unless the conditions of determination are exactly stated. The pH of a dry soil would be meaningless, since active hydrogen-ions can only exist in solution. Determinations of pH therefore refer to soil-water pastes or suspensions and, since different results can be obtained according to the soil-water ratio used, it is necessary to specify the actual ratio. Attempts have been made to determine pH under field moisture conditions, but the data usually encountered in the literature refer to

aqueous suspensions. It has been recommended by the International Society of Soil Science that a soil-water ratio of 1:2.5 should be used, but it should be clearly realised that, even where this recommendation is followed, the results are only of comparative value and are no absolute measure of a soil property.

It should be noted that the reaction of a suspension of a soil, even under standardised conditions of determination, is governed not only by the degree of saturation of the complex acids of the colloidal complex, but also by the nature of the complex itself. Broadly speaking, the greater the ratio of the acidoid groups, silicic acid and humic acid, to the basoid groups, aluminium oxide and ferric oxide, the more acid will be the reaction at a given degree of saturation. A lateritic soil may contain very small amounts of exchangeable bases and yet show a pH approaching neutrality, whilst a soil rich in organic matter or with a highly siliceous clay complex might show a pH of 4 or even less at the same degree of saturation.

In spite of these limitations, pH data do give an indication of the relative base status of similar soils and are of significance in connexion with many of the most important pedogenic processes. The lowest figures are found in peats and humus podsols, whilst the highest figures are found in alkaline soils. Red soils, brown forest soils, and black earths, occupy intermediate positions.

CONSTITUENTS SOLUBLE IN STRONG ACIDS.

Proceeding from the assumption that regulated extraction with strong mineral acids can separate the weathering complex of the soil from the unweathered portion, many investigators determine the amounts of silica, alumina, ferric oxide, and other bases soluble in hydrochloric or sulphuric acid under conventional conditions of working. In the writer's opinion, however, there is no evidence that any acid extraction can succeed in dissolving the weathering complex without attacking unweathered minerals in addition. Never-

theless, the results obtained by such methods may afford indications which may be of practical value as to the reserves of plant nutrients in the soil. Phosphoric acid is frequently determined in acid extracts, but no evidence is yet available that the amount thus brought into solution represents any definable category of the phosphorus compounds of the soil.

SOLUBLE SALTS.

In certain classes of soils, it is desirable to make determinations of the content of soluble salts. As a complete analysis of the soluble salts is generally a tedious operation, it is often considered sufficient to obtain an indirect estimate by determining the electrical conductivity of aqueous suspensions.

PHYSICAL DETERMINATIONS.

An account of soil analysis would be incomplete without some reference to the use of physical determinations for the description and definition of soils. But, apart from mechanical analysis, which is essentially a physical determination, it must be admitted that methods are still in process of evolution, and that no generally accepted system of physical determinations exists. It may be helpful, however, to review briefly those physical properties of the soil which, from the standpoint of soil definition, call for quantitative expression.

PORE SPACE AND VOLUME WEIGHT.

A knowledge of the variations in pore space and volume weight throughout the soil profile would doubtless be of considerable aid to the definition of the principal soil groups. But whilst the determination of pore space and weight per unit volume offer no difficulty with laboratory samples, determinations under field conditions are laborious and, except in almost stoneless soils, beset with considerable experimental difficulties. In spite of the importance of the information sought, it must be admitted that the determina-

tion of pore space has not yet become one of the routine investigations of the soil worker.

SINGLE VALUE DETERMINATIONS.

In recent years, attempts have been made to express those properties of the soil associated with their colloidal character by means of so-called "single-value" figures. Of those which may more properly be called physical determinations, hygroscopicity, moisture content at the "sticky point," and heat of wetting, have been most widely used.

Whilst the earlier workers on hygroscopicity attempted to determine full hygroscopicity in equilibrium with a saturated atmosphere, recent workers use atmospheres below saturation point. There is, at present, no uniformity of procedure. (See p. 173.)

The determination of the moisture content at the "sticky point" has, in recent years, attained considerable vogue among British workers, and since it does appear to have a definite physical meaning, i.e., the moisture content at which the soil gels are completely saturated, it should prove of great value in defining and describing soils. As it is determined with comparatively little labour and expense, it may be recommended for routine workers and will doubtless prove of increasing value as results are accumulated and examined in relation to other soil properties.

Heat of wetting has been used by certain American and continental workers, but has not come into general routine practice.

COLOUR DETERMINATION.

The colour of soils is an important aid in the recognition of the soil groups. Such groups as laterites, terra rossa, podsoles, and black earths, exhibit colours which are highly characteristic. These are almost invariably described in qualitative terms, with the result that, whilst a worker familiar with podsoles has a clear idea of the range of colours met with in these soils, his descriptions will not convey an

accurate impression to a worker familiar with tropical soils who may, indeed, use similar terms to describe unmistakably different colours.

The American soil survey workers have made efforts to define soil colours by methods such as the Munsell colour scale. Unfortunately, standards for comparison are neither convenient nor inexpensive, and no method of colour description has come into general use. The need for a routine method is very urgent, and it may be hoped that it will be met in the near future. (See p. 170.)

From the foregoing review of methods of soil analysis, it will be plainly realised that the instruments at our disposal are still very inadequate. This naturally follows from the incompleteness of our knowledge of the constitution of the soil and the novelty of much of the knowledge already acquired. Progress in the establishment of new methods of analysis is necessarily slow since, even when the investigator has decided what he wishes to determine, his methods must be tested on a wide variety of soils. A method may prove perfectly satisfactory for certain groups of soils and be utterly unsuited for others. It is for this reason that advances in methods of soil examination are now being sought by co-operative efforts among workers in different countries.

BIBLIOGRAPHY.

- ¹ KEEN, B. A., and COUTTS, J. R. H., " 'Single Value' oil properties : A study of the significance of certain soil constants." *J. Agric. Sci.*, 1928, **18**, pp. 740-765.
- ² WILLIAMS, RICE, "The determination of exchangeable bases in soils. Magnesium, potassium, and total bases." *J. Agric. Sci.*, 1929, **19**, pp. 589-599.

CHAPTER XIX.

SOILS, PLANT GROWTH, AND AGRICULTURE.

THE purpose of this work is mainly to present an account of the soil as a natural object of study. We have consciously refrained from presenting the problems of the soil from the standpoint of crop production, partly on account of the complexity of the considerations involved and partly because a satisfactory apprehension of the nature of the contract between the soil and the living plant must be preceded by a thorough understanding of the constitution of the soil. Yet an account of the soil as a natural object would be incomplete without some reference to its relationship to the higher plants which inhabit its surface. For this relationship not only constitutes the chief human interest in the soil, but is also of direct importance in the study of the development of the soil itself.

NUTRITION OF PLANTS.

The dry matter of plants consists of organic matter, and mineral or ash constituents. The principal constituents of organic matter are:—

Carbohydrates, including sugars and starches.

Fibres.

Oils, fats, and waxes.

Proteins.

The carbon, which forms about 55-60% of the dry organic matter, is obtained by the process of carbon assimilation or photosynthesis from the carbon dioxide which forms about 0.03% of the atmosphere. This change consists essentially in the formation of starch from carbon dioxide and water, probably through formaldehyde and dextrose as

intermediate stages, with concomitant liberation of oxygen. The reaction, or chain of reactions, is endothermic and the source of energy is the sunlight. A further condition for the process is the presence of chlorophyll, the green colouring matter of plants. Except in so far as carbon dioxide is produced in the soil by the oxidation of organic matter, the soil cannot be considered as of direct importance in carbon assimilation. The hydrogen and oxygen of organic matter are obtained ultimately from the soil moisture.

The remaining elements of the dry matter of plants, namely, nitrogen, and the elements present in the ash, namely, sulphur, phosphorus, calcium, magnesium, potassium, iron, silicon, and all the other elements which, though not actually essential to growth and reproduction, normally occur in plants, are obtained from the soil.

It is generally supposed that the elements thus obtained, often termed plant nutrients, enter the plant root by absorption from the soil solution; and much of the theory of plant nutrition has been built up on the basis of experiments in which plants are grown in so-called culture solutions. It is possible, however, that the mode of absorption by the plant may not be so simple as this theory implies. N. M. Comber¹ has put forward the suggestion that the colloidal material of the soil and the colloidal material of the plant roots may form a single system within which translocation of material may take place. Whatever be the nature of the contact between soil and plant, it is evident that a principal duty of the soil is to furnish the growing plant with an adequate and balanced supply of the nutrient elements.

There are, however, other factors involved in plant growth. Apart from the water required for the actual material of plants, water is required by the roots of plants to make good the constant losses by transpiration from the leaves. It has been found that, for every part by weight of dry matter formed, from 200 to 1,000 parts by weight of water are required. The supply of water to plants is intimately bound up with the supply of oxygen required for

the respiration of the root system of plants. An excess of water, whilst not necessarily disadvantageous in itself, restricts the air supply to roots. On the other hand, an excess of air, not in itself harmful, implies a deficiency of moisture. Between the two extremes lies the optimum, which is often considered to be fulfilled by a 50% saturation of the pore space of the soil with water.

The ease with which the soil yields up water to plants diminishes as the moisture content decreases below the optimum. Finally, a point is reached at which the water is so firmly held by the soil colloids that the losses by transpiration can no longer be met, and wilting, i.e., loss of turgidity, occurs. The moisture content at the *wilting point* is approximately 1.5 times the hygroscopic coefficient. It is thus evident that the moisture present in a soil is not entirely available to plants, but is subject to a deduction which is roughly dependent on the colloidalty of the soil.

Each plant has its own minimum, optimum, and maximum temperature for germination and also for adult growth. The value of any particular situation or soil for plant growth is therefore relative to the plants which can grow there. Plant growth cannot take place below 0°C , but certain arctic and alpine species can grow at temperatures immediately above freezing point. At the other extreme are plants which can tolerate temperatures up to $45\text{--}50^{\circ}\text{C}$. The optimum temperature for growth varies considerably among different plants.

Generally speaking, the temperature of the soil is closely related to that of the air; but, as we have seen in Chapter VII, minor differences may occur among soils even under the same climate. In regions where the growing season is limited by winter, such minor differences may be of considerable importance in their effect on the beginning and duration of the year's growth.

The next factor to be considered is negative, namely, the absence of injurious or toxic substances. The most obvious cases of inhibition of plant growth due to this group

of factors are seen in soils in the vicinity of lead, copper, tin, or zinc mines, where the presence of compounds of these metals partially or entirely excludes vegetation.

Another type of injury due to harmful constituents is that caused by the presence of excess of soluble salts, most commonly sodium chloride, sulphate, or carbonate. There is a considerable variation in the tolerance of different plants to soluble sodium salts. Indeed, some plants, natives of the sea shore and of saline soils, may be described as halophytic. Most economic plants, however, can only tolerate small concentrations of salts in the soil moisture, and the presence of such salts is a serious drawback to the agricultural use of soils in which they occur.

Finally, there is the possibility of injury or limitation of plant growth through the presence of acids, or, defining the case more strictly, the prevalence of an excessive concentration of hydrogen-ions in the soil moisture.

The problem of soil acidity in its relationship to plant growth has been for many years the subject of laborious investigation and active debate. That certain soils fall below satisfactory productiveness by reason of the prevalence of conditions which can be remedied by the application of dressings of lime or calcium carbonate cannot be doubted. That acidity in itself is the injurious factor cannot be so generally maintained. It is certainly true that an extreme degree of acidity, such as is rarely found in ordinary soils, is injurious to plant growth. It is also true that a reaction slightly on the acid side of neutrality inhibits the activity of the nitrogen-fixing bacterium *Azotobacter*, and in so far as this activity is a necessary part of the nitrogen economy of the soil, acidity is harmful. Otherwise, most plants can tolerate a fairly wide range of reaction, although the optimum reaction for individual species varies somewhat.

The problem is complicated by the fact that soil acidity cannot enter into the problem of soil fertility as an independent factor, for it is correlated with a deficiency of available calcium and also, in some cases, with the presence of actively

toxic aluminium compounds. Further, in some cases, soil acidity may be correlated with undesirable physical conditions. It is thus not easy to assert generally that soil acidity, where present, is a limiting factor. The limitation may actually be due to lack of available calcium, the presence of active aluminium, or the prevalence of an undesirable physical condition.

SOIL FERTILITY.

We have now considered the factors concerned in the growth of plants, and it is evident that fertility depends on a due balance of all these factors. As shown by J. von Liebig, plant growth is limited by that factor which operates at a minimum. We may illustrate fertility by the analogy of a chain, whose strength is that of its weakest link. The weakest link is the limiting or minimum factor and the first step in the improvement or raising of fertility is the improvement of the limiting factor. Thus, if deficient water supply be the limiting factor, then an improvement in this respect is the first step in increasing plant growth. The principal task of applied pedology is the discovery of limiting factors and the means of their improvement.

Fertility can only be postulated in terms of particular crops. Thus, certain sandy soils are fertile for market garden crops, but unsatisfactory for grass. On the other hand, the Lias clays of the Eastern Midlands of England are fertile as pasture grounds, but unsatisfactory for market garden crops. Soil fertility might be defined in general terms as the ability of a given soil to grow satisfactorily some or all of those crops which are permitted by the regional climate. In most habitable lands, fertility and infertility depend mainly on the moisture and air relationship of soils. The most frequent cause of infertility is deficiency or excess of moisture.

SOILS AND AGRICULTURE.

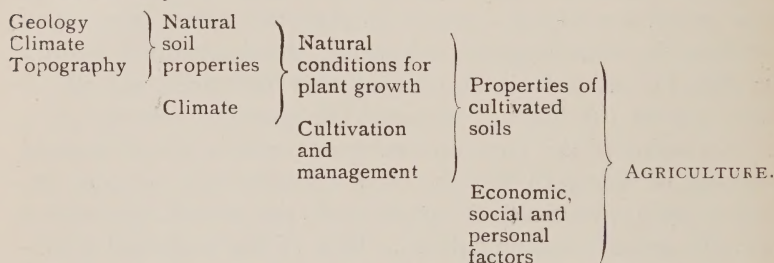
The earliest cultivations of soil were in arid and semi-arid regions, and it has been shown that in the westward

migrations into Europe the first settlements were in forest-free regions, the loess areas, which in their general character resemble steppe. For primitive agriculture, the steppe or prairie offers a much better prospect for successful crop production than the forest soils of humid regions. Under steppe, the climatic conditions, though drier than in humid regions, are more dependable. There is generally sufficient moisture after the winter for the establishment of crops, whilst the hot dry summer is favourable to ripening and harvest. Further, the moderate leaching tends to the conservation of plant nutrients and the inhibition of the deteriorative processes which operate in humid soils. The chief problem is water supply and yields are determined mainly by this factor. It is significant that the principal wheat areas are on the black earths and the chestnut earths.

In spite of the more intensive agriculture of the humid regions of Western Europe, the cultivator is actually engaged in a perpetual struggle with the conditions which militate against arable culture. One of the principal differences between soils of the black earth group and soils of the podsollic group is in their structure. Whilst the calcium-saturated soils of the former class naturally tend towards the desirable crumb structure and demand a minimum of cultivation, the podsollic soils tend to assume the single-grain structure. The cultivator is therefore always attempting to establish a structure which is purely artificial. The leaching consequent on the high rainfall results in an impoverishment in mineral constituents, principally lime, which, if unchecked by applications of calcareous dressings, eventually results in deterioration, accompanied by loss of the desirable crumb structure. The actual labour of husbandry is increased by the fight against weeds, and the anxieties of harvest operations. High yields can be obtained in arable farming in the humid districts of western Britain; but it should be realised that, since the natural vegetation is forest or heath, constant effort is necessary to prevent deterioration. It is perhaps no exaggeration to say that the efforts

of the farmer in the humid regions are directed towards the production and maintenance of artificial black earths.

The agriculture of a particular locality is the reflection of a complex of factors, of which climate, soil, and situation are the most important, since they impose certain limits on the choice of crops for cultivation. But economic, social, and even personal factors also intervene, with the result that the agriculture of a country only partially reflects the variations in soil and climate within its borders. The interrelationships of these factors may be thus represented schematically :—



SOIL DETERIORATION.

In a period of agricultural depression, when the tendency is towards restriction of effort, deterioration is likely to ensue, and to be most pronounced in those regions where conditions are most artificial. The consequences of "low farming" are likely to be less serious where conditions approach more closely to those of the steppe and the prairie.

The tendency to degeneration of soils in humid temperate regions is always present, but the highly developed agriculture of countries such as England, Holland, and Belgium, is sufficient proof that the struggle against it is no impossible task and that, with cheap fertilisers and skilful tillage, crop yields can be obtained which would not be possible in drier regions.

Under the conditions of the humid tropics, the maintenance of an artificial system of land utilisation is beset with even greater difficulties. The tendency to deterioration of

virgin land brought into cultivation is far more rapid than in temperate climates. The high temperatures lead to a very rapid destruction of reserves of organic matter with consequent deterioration in physical conditions, whilst the torrential tropical rains, acting on such soils, produce vigorous erosion. It is not surprising, therefore, that there are many instances of newly broken soils becoming completely derelict within a few years through injudicious management.

Cultivated soils are, to a large extent, artificial and show profiles which may differ markedly from the corresponding virgin profile. From the moment that the original cover of natural vegetation is destroyed, the processes which have produced the original succession of horizons are modified, and new factors, leading to a new profile succession, are introduced. Loss of organic matter is the commonest consequence of cultivation and may be demonstrated under practically all climates when pioneer cultivation is instituted. F. T. Shutt² gives an instance of a Manitoba prairie soil which showed a decrease of nitrogen content from 0.651% to 0.506%, and of ignition loss from 19.43% to 14.79%, as the result of 25 years of cultivation. The writer found a cultivated soil in Caernarvonshire to contain 0.490% nitrogen, and to have an ignition loss of 16.6%, whilst the adjoining waste, from which it had been reclaimed, contained 1.106% nitrogen and gave 43.1% ignition loss.

Loss of organic matter is not the invariable consequence of bringing land into cultivation. Where grass husbandry is practised, or where heavy dressings of organic manures are given, as in market gardens, the organic matter content of the soil may be maintained or even increased. In general, the loss of organic matter is least where livestock farming is practised, for it is then possible to return to the soil a considerable proportion of the crops raised and fed to stock, in the form of manure.

In many parts of the world, the most serious consequence of cultivation is erosion. We have already referred to the widespread erosion in the Eastern United States,

which has taken place within comparatively recent times. Through uninstructed utilisation of land in past generations, much irreparable damage has been done over large areas of the tropics and sub-tropics. It is important, therefore, that future agricultural operations should be conducted with the fullest possible precautions against further losses.

The crops obtained immediately after bringing virgin land under cultivation are always bigger than succeeding crops. It is the common experience that yields fall off as the land is longer under cultivation. Under virgin conditions, the demands on plant nutrients are comparatively slight, for a large proportion of the nutrients extracted from the soil by plant roots is returned to the soil again in the form of plant residues. Under pioneer cultivation, the soil is cropped continuously and deterioration is inevitable. Firstly, there is a rapid exhaustion of plant nutrients as crop after crop is removed, and secondly, there is, in many cases, a physical deterioration in the soil consequent on loss of organic matter. This is manifested by a decreased water holding capacity and an increased liability to erosion.

The deterioration consequent on pioneer cultivation of virgin land may be checked where live stock husbandry is practised, for it is then possible to return to the soil a considerable proportion of the nutrients extracted from it by crops. Organic matter may also be maintained by green manuring. The introduction of artificial fertilisers, whereby the plant nutrient status of the soil can be controlled, marks a further stage in land utilisation. With the present abundance and cheapness of fertilisers, the serious problems of continuous agriculture are mainly physical in character. Indeed, in some systems of agriculture, crops are grown almost entirely on added fertilisers, and the rôle of the soil is mainly physical. This is the case over a large part of the cotton lands of the United States.

Even where the land is not brought under actual cultivation, uncontrolled grazing may lead to deterioration of the soil. Examples of this may be cited from South Africa,

where injudicious grazing and close folding (kraaling) has led to a destruction of the natural cover. The consequence of this has been erosion by gullying, resulting in lowering of the water-table, and accentuation of the effects of drought.

The utilisation of land for irrigation brings its own set of problems. In many instances, the continued addition of irrigation water has led to a rise in the water-table, whereby injurious salts have been brought to the surface. In other cases the irrigation waters used contain excess of sodium over calcium salts, with the result that the irrigated soil is gradually changed from a calcium to a sodium soil with consequent deterioration and ultimate disaster to crops.

CONCLUSION.

The deterioration of soils under cultivation, though only operative in certain regions, may be, in the future, a serious menace to the world supply of food and other commodities obtained from the soil. Much of the damage done in past generations is well-nigh irreparable, and the prevention of further losses is one of the main tasks confronting the pedologist. Hitherto, the practical aims of soil investigation have been chiefly concerned with the increase of crop growth. In future, the problems of soil conservation must claim more attention, particularly in view of the potential developments of tropical agriculture. These problems demand the most ample elucidation of the constitution and properties of the soil. The successes which have been obtained encourage the hope that the fullest utilisation of the world's soil resources may not be incompatible with the conservation of those resources by systems of permanent agriculture. *There shall be an handful of corn in the earth upon the tops of the mountains* (Ps. lxxii, 16).

BIBLIOGRAPHY.

- ¹ COMBER, N. M., "The availability of plant food: A modification of the present hypothesis." *J. Agric. Sci.*, 1922, **12**, pp. 363-369.
- ² SHUTT, F. T., "The influence of grain growing on the nitrogen and organic matter content of the western prairies soils of Canada." *J. Agric. Sci.*, 1925, **15**, pp. 162-177.

APPENDIX.

METHODS OF ANALYSIS.

THE methods to be described in the present appendix are principally such determinations as may serve for the characterisation of a soil. They are routine methods, in which refinement is to some extent subordinated to convenience. Such is the innate variability of soils that available effort is usually more economically expended in securing data for a wide range of soils with a moderate degree of accuracy, than in practising the utmost refinements of analysis on a limited number of soils. For certain purposes, however, refined methods are required. Such methods are necessary in the intensive study of the constitution and properties of individual types, or the changes in individual soils under cultivation or degeneration.

These considerations are exemplified by the case of mechanical analysis. In a study of the probable error of sampling in survey work, G. W. Robinson and W. E. Lloyd (*J. Agric. Sci.*, 1915, **7**, pp. 144-153) found that the probable error due to variation in an apparently uniform field ranged from 2.4 to 12.4% for the individual determinations in a mechanical analysis, the highest figure being for the fine gravel determination. Excluding this variable constituent, the average field error was about 3.5%. The average probable error of sampling from the laboratory sample is rather less. For ordinary routine purposes, it would not appear necessary to aim at an accuracy of determination much greater than 3-4%. Such a degree of accuracy would ensure that the final result, affected also by field variations, would have an accuracy of about 5%.

If the figure given by a method of analysis for the clay content of a soil be 25%, it would be sufficiently accurate for

descriptive purposes if this is within 1% (i.e., 4% of the clay) of the true value. On the other hand, to take a hypothetical case, if it were desired to study the rate of chemical weathering as shown by the clay content, then a method would be required which would be capable of showing up variations such as might be expected within the period of the experiment. Additional accuracy might be attained by replication of analyses and by more careful standardisation of methods.

The degree of accuracy to be sought in any series of determinations must always be considered in connexion with the scope and purpose of the work. For descriptive purposes in connexion with soil surveys, it is generally necessary to sacrifice accuracy to some extent in order to obtain data for as wide a range of soils as possible.

SAMPLING.

For the purpose of analytical determinations which have as their aim the assessment of the manurial requirements of the soil of a field, it is usual to work with composite samples. For example, cores of soil drawn by a cylindrical auger from 6-10 places distributed over a field are mixed together to give a composite representative sample of the field. This method is certainly to be recommended if it is desired to form an estimate of lime or phosphate status for manurial recommendations.

For purely scientific purposes, however, the composite sample should be avoided. A trench or pit should be dug to a sufficient depth to expose all the soil horizons. Samples should then be taken from each recognisable horizon, ensuring that the same amount of material is taken from each depth. Where, as in certain alluvial and æolian soils, no definite horizons are recognisable, samples may be taken of successive layers of 6 in. or 12 in.

PREPARATION OF SAMPLE FOR ANALYSIS.

The sample is allowed to dry on a plate maintained at about 40°C. It is then broken down with a wooden pestle and passed through a 2 mm. sieve, the material remaining on the sieve being reckoned as stones.

The 2 mm. sample is used for analysis. For certain purposes, notably for determinations of organic carbon and nitrogen, a finer sample is prepared by grinding a sub-sample completely through a 100 mesh sieve.

MECHANICAL ANALYSIS.*

The method recommended is a modification of that prescribed by the International Society of Soil Science. Ten grams of soil are heated with 60 c.c. of 6% hydrogen peroxide, with further additions of 50 c.c. portions until vigorous frothing due to oxidation of organic matter ceases (A). In some cases the presence of manganese dioxide causes such a rapid decomposition of the peroxide that the attack on the organic matter is hindered. In such cases, the soil is first digested with 0.2-0.5 g. sodium hydrogen sulphite and 20 c.c. water until nearly all the water has evaporated. The oxidation can then be proceeded with.

When oxidation is completed (B), sufficient N-hydrochloric acid and water are added to give 100 c.c. of 0.2 N-hydrochloric acid. Where gypsum, or carbonates in amounts greater than about 5% are present, excess acid should be added (C). After standing for one hour with frequent shaking, the contents of the beaker are filtered. It is convenient to carry out this process, aided by suction, on Buchner funnels to which 9 cm. hardened filter papers have been attached by cellulose acetate (Durofix) cement. Washing is carried out with three 50 c.c. lots of distilled water. The washed residue is transferred by a jet of hot water to a 70 mesh sieve and pestled with a rubber pestle until a clean residue is left on the sieve. This is collected, dried at 105°C, and weighed to give the coarse sand (2 mm.-0.2 mm.). The turbid liquid which has passed the 70 mesh sieve is transferred to a pint milk bottle, 4 c.c. of N-NaOH are added (D), and the bottle is shaken for two hours on a reciprocating shaker, or for 6 hours on an end-over-end shaker.

*The methods included in this appendix will not, in general, be described in detail. An exception is made in the case of mechanical analysis in view of its importance for characterising soils and, also, because, no full account of the improved procedure is generally accessible.

The volume of suspension to be shaken should be 300-400 c.c. After shaking, the suspension is transferred to a 500 c.c. cylinder and made up to the mark. It is then shaken for one minute by hand with repeated inversion and allowed to stand for 4 mins. 48 secs. at 20°C. The settling times for other temperatures are given in Table A.

A 20 c.c. pipette, fixed through a wide cork, is so adjusted that, when the cork rests on the top of the cylinder, the point is 10 cm. below the surface (E). At the conclusion of the settling period (F) appropriate for the temperature, the pipette is carefully lowered into position, taking care to prevent ingress of liquid until lowering is complete. A 20 c.c. sample is withdrawn and is delivered into a tared flat porcelain dish, about 9 cm. diameter, which is then dried at 105°C and weighed. The weight of dry material in milligrams divided by 4 gives the percentage of material finer than 0.02 mm. (silt and clay). The figure after ignition may also be determined if desired.

The suspension is again thoroughly shaken and allowed to settle for the clay sampling, which is carried out at 10 cm. below the surface, after settling for the appropriate time. The weight of material, dry at 105°C, divided by 4 gives the percentage of material less than 0.002 mm. (clay). This figure, subtracted from the clay and silt, obtained in the previous sampling, gives the percentage of silt.

TABLE A.—SETTLING TIME FOR SILT AND CLAY AT DIFFERENT TEMPERATURES.

Temperature	Settling times for sampling at 10 cm. depth			
	Silt		Clay	
Deg. C.	Mins.	Secs.	Hrs.	Mins.
10	6	14	10	24
15	5	27	9	5
20	4	48	8	0
25	4	19	7	7
30	3	44	6	22

Sampling may be carried out at different depths if the time is proportionately altered, e.g., 20 cm., and 16 hours for the clay.

The greater part of the supernatant suspension in the cylinder is now poured away and the sediment at the bottom is washed into a beaker, well stirred, and made up to 10 cm., allowing for the thickness of sediment at the bottom. After standing for the time indicated for the silt fraction (4 min. 48 secs. at 20°C), the supernatant liquid is poured away. The sediment is again stirred and made up to 10 cm., allowed to stand and decanted as before, the operation being repeated until no material remains in suspension at the end of the period. The residue is the fine sand (0.2-0.02 mm.).

In the filtrate after the peroxide and acid treatment the dissolved sesquioxides are determined after precipitation with ammonium chloride and ammonia, collection, washing, and ignition. The percentage found is added to the clay, from which it has been dissolved.

The percentage of moisture in the original soil is determined by drying 5 g. to constant weight at 105°C. It is convenient to combine this with a determination of the loss on ignition.

Carbonates (reckoned as calcium carbonate) may be determined by means of the Collins calcimeter (*J. Soc. Chem. Ind.*, 1906, **25**, p. 518), or by the method of H. B. Hutchinson and K. McLennan (*J. Agric. Sci.*, 1914, **6**, pp. 323-327).

The results are set out as follows:—

Coarse Sand (2 mm.-0.2 mm.)	
Fine Sand (0.2 mm.-0.02 mm.)	
Silt (0.02 mm.-0.002 mm.)	
*Clay (<0.002 mm.)	
Moisture	
Calcium Carbonate	
Organic matter decomposed by H ₂ O ₂ , etc.	
Total	100.0
*Including loss by solution	

NOTES ON MECHANICAL ANALYSIS.

A. In the case of soils containing less than 2% of organic matter, the omission of the peroxide treatment is without serious effect on the efficiency of dispersion. For soils containing up to 2-3% organic carbon about 50 c.c. H_2O_2 is usually sufficient.

E. K. Troell (*J. Agric. Sci.*, 1931, **21**, pp. 476-483) recommends the substitution of sodium hypobromite for peroxide as an oxidising agent for organic matter. In this method, no acid treatment is introduced, and therefore, with soil containing carbonates, the results are not strictly comparable with those by the H_2O_2 -HCl-NaOH method. With carbonate-free soils, however, it appears to be a satisfactory alternative, and may be useful in tropical climates where difficulty is experienced with obtaining and keeping 6% hydrogen peroxide. It should be added, however, that satisfactory dispersion can be obtained with most soils, even with 2% hydrogen peroxide, if larger volumes are used.

B. With carbonate-free soils, it is more convenient to pass the material through the 70 mesh sieve before acid treatment. There is then no necessity for pestling and the residue in the sieve can be obtained clean by simple washing. With soils containing carbonate, this can also be done if the residue on the 70 mesh sieve is given an acid treatment to dissolve carbonates. With soils containing coarse gypsum it is always preferable to sieve after peroxide, as it reduces the amount of gypsum to be removed by acid treatment.

C. In the case of gypseous soils, the coarse gypsum will have been previously removed by sieving, as recommended above. If 0.5 N-acid is used instead of 0.2 N, and shaking is included, the fine gypsum can usually be completely dissolved; but washing should be prolonged until the disappearance of sulphate reaction in the washings. Exceedingly gypseous soils may require still more hydrochloric acid. Fine gypsum may also be removed by washing with 10% ammonium acetate solution.

D. The method of A. N. Puri (*Mem. Dept. Agric. India, Chem. Series*, 1929, **10**, No. 8), in which the soil is leached with *N*-sodium chloride to obtain the sodium soil and suspended in water after removal of excess of sodium chloride by washing, adding, if necessary, enough *N*-sodium hydroxide to give an alkaline suspension, gives excellent results with soils low in organic matter. Strict comparison with the H_2O_2 -HCl-NaOH method is, however, only possible with carbonate-free soils. With soils containing much organic matter the method may be modified by introducing a preliminary peroxide treatment.

A large number of comparisons made over a wide range of soils in the writer's laboratory by Dr. Minnie Richardson has shown that the substitution of sodium hydroxide for ammonia, suggested by the Puri method, gives comparable results for all soils which can be satisfactorily dispersed by ammonia, as recommended in the International method, and gives complete dispersion in those cases where ammonia is unsatisfactory. The rapidity of dispersion with soda is such that satisfactory results can be obtained with most soils by shaking vigorously by hand for a few minutes.

E. Various types of apparatus for sampling are now on the market. Whilst satisfactory results can be obtained by hand manipulation of an ordinary pipette, we have used a pipette with a three way cock so arranged that in one position the pipette may be filled, using pump suction, whilst in the other position it delivers exactly 20 c.c.

F. The cylinder should stand at a fairly constant temperature away from any sources of heat in order to avoid disturbance by convection currents.

HYGROSCOPIC MOISTURE.

The hygroscopic moisture present in the air-dry soil may be obtained from the loss in weight on drying to constancy at 100-105°C. The figure obtained is in itself of slight significance on account of the varying humidity of the

atmosphere, but it is required in order to calculate other analytical results to the basis of oven-dry soil.

The hygroscopic capacity is obtained by allowing soil to attain equilibrium with an atmosphere of standard humidity. A convenient procedure is that of A. N. Puri (*J. Agric. Sci.*, 1925, **15**, pp. 273-283), in which the soil is allowed to reach equilibrium with an atmosphere of 50% humidity by being kept in a vacuum desiccator over sulphuric acid of density 1.330 (23.4 c.c. sulphuric acid to 57 c.c. water).

LOSS ON IGNITION.

This determination, though untrustworthy as a measure of organic matter, may give a rough indication of the colloidal nature of a soil. It is obtained from the loss on igniting oven-dry soil to constant weight. If calcium carbonate is present there is a certain loss by carbon dioxide. By moistening the ignited material with ammonium carbonate solution and igniting gently, calcium carbonate is re-formed and the loss in weight is due simply to organic matter and combined water.

ORGANIC CARBON AND NITROGEN.

Organic carbon is determined to obtain a measure of the organic matter content of the soil. For ordinary purposes the organic carbon multiplied by the factor 1.724 may be taken as equivalent to the organic matter.

Organic carbon and nitrogen may be determined by the method of G. W. Robinson, W. McLean, and R. Williams (*J. Agric. Sci.*, 1929, **19**, pp. 315-324). The method is essentially based on the Kjeldahl reaction. The sulphur dioxide from the oxidation of the organic matter is collected in standard iodine solution in an absorption tower. Comparisons with results of dry combustion examinations indicate that the figures obtained by this method should be corrected by the factor 1.116. Nitrogen is determined in the contents of the Kjeldahl flask by distillation with excess of 50% NaOH, into 0.02 N- H_2SO_4 , using methyl red as an indicator and heating at the end point.

G. W. Robinson and W. McLean (*J. Agric. Sci.*, 1930, **20**, pp. 345-347) have shown that certain soils contain appreciable proportions of coal and coke. A correction may be applied by oxidising, decanting repeatedly at 10 cm. after 250 secs. and determining the carbon in the residue, which contains about 75% of the coal, etc., originally present.

COMPOSITION OF CLAY FRACTION.

If the ordinary dispersion treatment be used for obtaining the clay fraction, some inaccuracy is introduced owing to the loss of sesquioxides in solution. As a provisional method, peroxide treatment followed by leaching with sodium chloride, as in the Puri method, may be recommended. The clay is then removed by repeated sedimentation. The combined pourings off are flocculated by calcium chloride, collected, dried, ignited, and ground to pass through a 100 mesh sieve. Silica, alumina, ferric oxide, titania, and, if desired, the alkalis and alkaline earths are determined by the usual methods of silicate analysis. If alkaline earths are determined, flocculation with calcium chloride should, of course, be omitted, and the united pourings off obtained by evaporation.

NOTE. — In the writer's opinion, the clay fraction, as recognised by the International Scale, may be too coarse for this purpose, and the use of a time/depth ratio of 8.6 cm./24 hours is to be preferred. The use of an even smaller settling velocity, such as 8.6 cm./10 days ($\log V = 5.0$) would render the exclusion of unweathered minerals even more certain.

EXCHANGEABLE BASES.

These may be determined by the method of D. J. Hissink (*Int. Mitt. Bodenkunde*, 1922, **12**, pp. 81-172) in which 25 g. of soil are leached with N- sodium chloride solution. For carbonate-free soils, leaching with 0.5 N- acetic acid as described by Rice Williams (*J. Agric. Sci.*, 1928, **18**, pp. 439-445; 1929, **19**, pp. 589-599; 1930, **20**, pp.

355-358) may be recommended. Ammoniacal nitrogen may be determined by the method of W. McLean and G. W. Robinson (*J. Agric. Sci.*, 1924, **14**, pp. 548-554).

For soils containing carbonates, the method commonly employed is that suggested by D. J. Hissink in which the soil is leached with N- sodium chloride solution. The first and second litres of the extract are analysed and the amount of calcium in the second litre is taken as a measure of the solubility of the calcium carbonate in the first. E. M. Crowther and J. K. Basu (*J. Agric. Sci.*, 1931, **21**, pp. 689-715) have pointed out that this method is subject to errors caused by the presence of carbon dioxide, which increases the amount of calcium brought into solution. They suggest a modification by which the true amount of calcium brought into solution as carbonates and bicarbonates is determined by titration to a methyl-red end point in boiling solution. The exchangeable calcium is found by subtracting the carbonate and bicarbonate content from the total calcium content of the extract.

SOIL REACTION.

This may be determined by the quinhydrone electrode method of E. Biilman (*J. Agric. Sci.*, 1924, **14**, pp. 232-239). For more accurate work the hydrogen electrode method as developed by E. M. Crowther (*J. Agric. Sci.*, 1925, **15**, pp. 201-221) may be used.

S. G. Heintze and E. M. Crowther (*Trans. Ind Comm. Int. Soc. Soil Sci.*, 1929, A., pp. 102-111) have discussed the errors introduced into the quinhydrone method by the presence of manganese dioxide in soils. For a full discussion of the problem, the report of a committee of the International Society of Soil Science (*Soil Research*, 1930, **2**, pp. 77-140) may be consulted. Where the presence of manganese dioxide is suspected, the potentiometer reading should be taken as soon as possible (10 secs.) after addition of quinhydrone, as a rapid drift in potential resulting in higher values for pH occurs. Even with this precaution, the results obtained with manganimiferous soils are higher than with the

standard hydrogen electrode. The quinhydrone method is inapplicable above pH 8.3.

For approximate purposes, the reaction of the soil may be determined colorimetrically in soil extracts, but the results should be checked against electrometric determinations. A mixed indicator (e.g. B.D.H. soil indicator) is convenient for approximate determinations of soil reaction in the field.

MOISTURE CONSTANTS.

B. A. Keen and H. Raczkowski (*J. Agric. Sci.*, 1921, **11**, pp. 441-449) have devised a simple method, improved by J. R. H. Coutts (*J. Agri. Sci.*, 1930, **20**, pp. 407-413) for determining saturation capacity, whereby soil is placed in a brass box with a perforated bottom in contact with water. From a simple series of weighings, it is possible to determine: (1) apparent density; (2) real density; (3) pore space; (4) saturation capacity per unit volume of soil; and (5) the volume expansion on saturation. For details, the original paper should be consulted.

The moisture content at "sticky point" advocated by F. Hardy (*J. Agri. Sci.*, 1923, **13**, pp. 243-264) is determined by mixing soil with varying quantities of water until it can be just kneaded without showing external adhesion, i.e., without soiling the hands or adhering to the instrument with which it is worked. B. A. Keen has used an incorporating machine for this purpose. When the "sticky point" is reached, a small sample is taken and its moisture content determined.

Moisture equivalent determinations require expensive apparatus, but a value related to this constant may be obtained by the method of G. Bouyoucos (*Soil Sci.*, 1929, **27**, pp. 233-240), in which a quantity of soil is placed on a Buchner funnel, thoroughly washed, and given a one inch irrigation with water. The funnel is connected to a vacuum pump and drainage is allowed to proceed for ten minutes after standing water has disappeared from the surface. The moisture content is then determined.

PLASTIC LIMITS.

An approximate indication of plasticity may be obtained by determining the upper and lower limits of plasticity by the method of A. Atterberg (*Int. Mitt. Bodenkunde*, 1911, **1**, pp. 10-43). In the determination of the upper limit, water is mixed with the soil until a definitely fluid paste is obtained. Dry soil is then added until, with thorough mixing, the upper limit of plasticity is reached. This is tested by placing the paste in a round porcelain dish (about 10 cm. diameter), cutting a V-shaped furrow and tapping the bottom of the dish smartly on the palm of the hand. So long as the paste is fluid the furrow closes up. The upper limit of plasticity is reached when the paste no longer runs together after being furrowed and tapped. A small sample is then taken and its moisture content determined.

To determine the lower limit, more dry soil is added and the paste is moulded into cylindrical form and rolled on paper with a flat strip of wood. The limit is reached when the cylindrical lumps crumble to pieces on rolling, and the moisture content is then determined.

The plasticity number is the difference between the moisture content at the upper and that at the lower limit of plasticity, the figures being expressed on the dry soil in each case.

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